Bulletin
of the
Geological Society of America

VOLUME 52 MAY 1, 1941 NUMBER 5

CONTENTS

	PAGES
Geology of the zinc-lead deposit on Calumet Island, Quebec. By W. W. Moorhouse	601–632
Seismic studies of floored intrusives in western New Hampshire. By F. C. Kruger and D. Linehan, S. J.	633–648
Roots of the Taconic fault in west-central Vermont. By H. E. Hawkes, Jr.	649–666
Mylonization of hybrid rocks near Philadelphia, Pennsylvania. By Elizabeth Armstrong	667–694
Geology of the Stony Mountain stock, San Juan Mountains, Colorado. By McClelland Dings	695–720
Changes in sea level, postglacial uplift, and mobility of the earth's interior. By Beno Gutenberg	721–772

Subscription, $10 per year.

Publication Office: Florida Avenue and Eckington Place, Washington, D. C.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Washington, D. C., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PRESS OF JUDD & DETWEILER, INC., WASHINGTON, D. C.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

Shore benches of North Pacific Islands. By H. T. Stearns.

Late Tertiary geology of Mono Lake, California. By Charles M. Gilbert.

Structural features of the Virgin Spring area, Death Valley, California. By Levi F. Noble.

Structure and petrology of the Boxelder laccolith, Bearpaw Mountains, Montana. By William T. Pecora.

Deformation in the interval Mt. Lyell–Mt. Whitney, California. By Evans B. Mayo.

Structure and metamorphism in the Mount Washington area, New Hampshire. By Marland P. Billings.

Cambrian geography and sedimentation in the central Cordilleran region. By Charles Deiss.

Silurian lithology in western Tennessee and adjacent states. By John R. Ball.

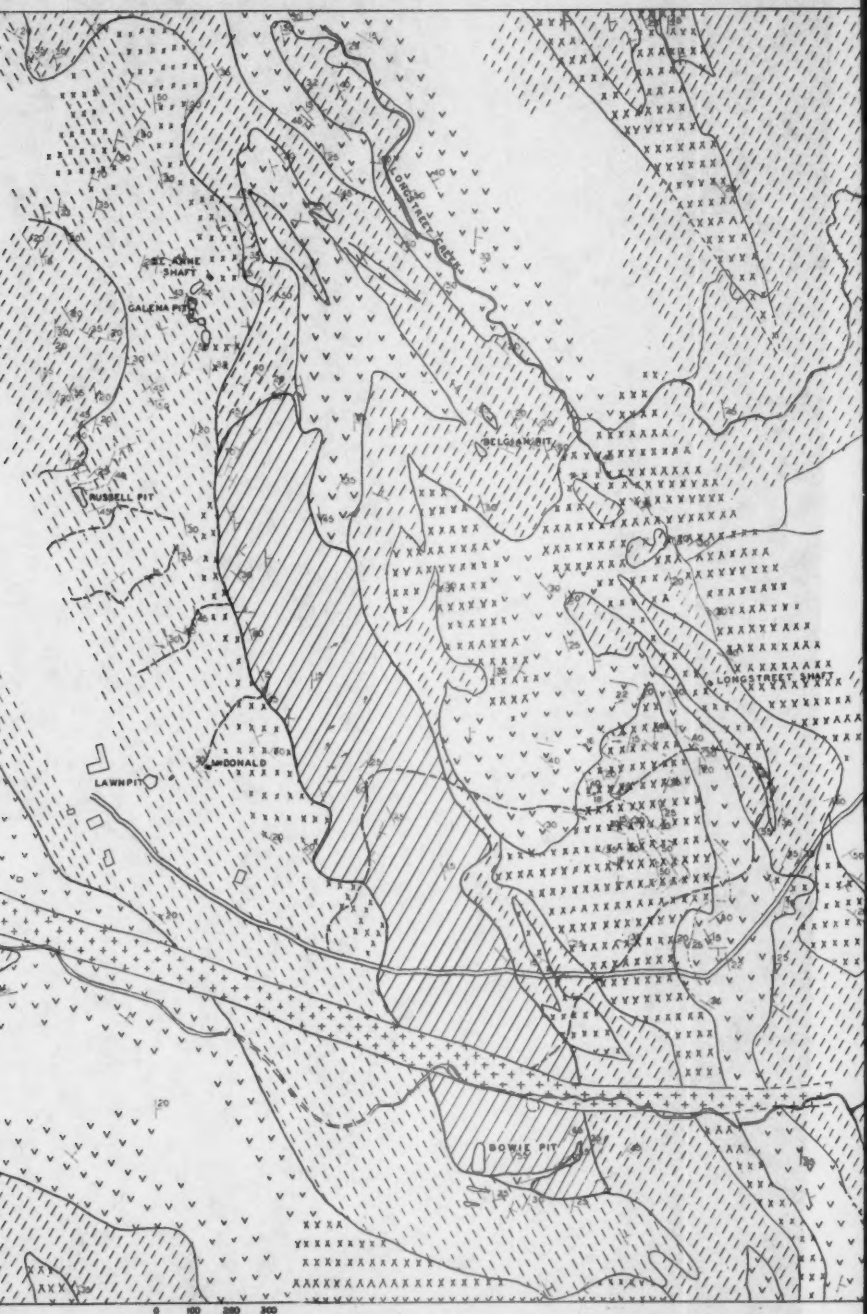
Early Devonian and late Silurian formations of southeastern Pennsylvania. By C. K. Swartz and F. M. Swartz.

Transportation of rock by kelp. By K. O. Emery and R. H. Tschudy.

Multiple glaciation of Mauna Kea, Hawaii. By C. K. Wentworth and William E. Powers.

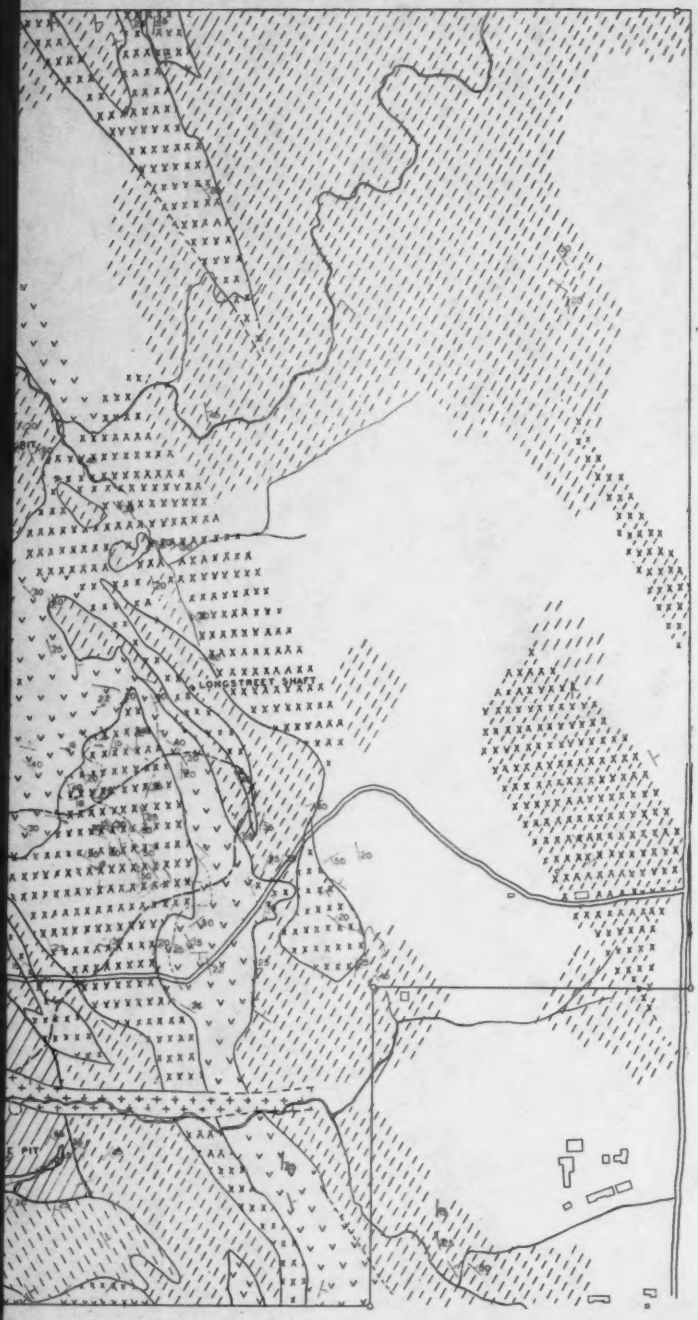
Mammalian fauna from the Pliocene of Honduras. By E. C. Olson and P. O. McGrew.

GEOLOGIC



GEOLOGIC MAP OF VICINITY OF CALUMET ISLAND ZINC-LEAD DEPOSIT

Topographic base prepared by Gordon Wells for Calumet Mines



LEGEND

DRIFT AREAS

CRYSTALLINE LIMESTONE AND DOLOMITE

AMPHIBOLITE WITH CARBONATE

AMPHIBOLITE AND AMPHIBOLITE GNEISS

"RUSTY-WEATHERING" GNEISS

GRANITIZED "RUSTY-WEATHERING" GNEISS

MEDIUM MAFIC GRANITIZED AND PARGASITE GNEISS

LEUCOCRATIC GRANITIZED GNEISS

DIABASE DIKE

SYMBOLS

BUILDINGS

PITS AND TRENCHES

SHAFT

ROAD

CREEK, SURVEYED AND UNSURVEYED (BROKEN LINES)

APPROXIMATE BOUNDARIES OF ORE BODIES (PROJECTED)

GEOLOGICAL BOUNDARY

GEOLOGICAL BOUNDARY INFERRED

FOLIATION (BEDDING ALSO IN PART)

VERTICAL FOLIATION

RELICT STRUCTURE

GENERALIZED STRIKE OF BEDDING

DIRECTION AND PITCH OF LINEATION

D ZINC-LEAD DEPOSIT

lumet Mines

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 601-632, 4 PLS., 3 FIGS. MAY 1, 1941

GEOLOGY OF THE ZINC-LEAD DEPOSIT ON CALUMET ISLAND, QUEBEC

BY W. W. MOORHOUSE

CONTENTS

	Page
Abstract	602
Introduction	602
Acknowledgments	603
Location, history, topography	603
General geology and structure	605
General considerations	605
Distribution and structure of lithologic units	605
General structural features	608
Petrology	609
Introduction	609
Carbonates	610
Amphibolite	611
"Rusty-weathering" gneisses	613
Granitized gneisses or migmatites	615
Mineralization	617
The ore zone	617
Paragenesis of the ores	620
General statement	620
Pyrite	621
Arsenopyrite	621
Sphalerite	621
Pyrrhotite	622
Galena	622
Chalcopyrite	622
Marcasite	622
Relations of sulphides and gangue	623
Physical conditions of mineralization	624
Nature of mineralizing solutions	624
Source of pegmatite and mineralizing fluids	626
General summary	626
Type of ore deposit	627
Works to which reference is made	627

ILLUSTRATIONS

Figure	Page
1. Location map	604
2. Longitudinal and cross sections of the Bowie ore body	618
3. Contours on top and bottom of the three known ore bodies, measured from an arbitrary datum	619

Plate Facing page
1. Geologic map of vicinity of Calumet Island zinc-lead deposit 601
2. Photomicrographs of amphibolites and gneisses 630
3. Gneiss and paragenesis of ores 631
4. Paragenesis of ores 632

ABSTRACT

The zinc-lead deposit described is located on Calumet Island, Quebec, 53 miles northwest of Ottawa. The complex ores resemble those of Edwards, New York, and Montauban, Quebec, and other Grenville deposits. They occur in moderately inclined Grenville metamorphic rocks, including crystalline limestone, amphibolite, amphibolite-gneisses, quartzitic (rusty-weathering) gneisses, and various granitized gneisses (migmatites). These represent argillaceous and calcareous sediments, contact metamorphosed and injected by a granitic magma of relatively high fluidity.

The sulphide minerals comprise pyrite, sphalerite, pyrrhotite, galena, tetrahedrite, chalcopyrite, and marcasite, deposited in the order named. The sulphides form lean disseminations in all of the rock types in the ore zones, and lenticular masses largely deposited in carbonate and contact metamorphic rocks. They replace carbonate, quartz, and all silicates except serpentine. The association of the ore minerals with quartz and contact silicates indicates that they accompanied the same agents that produced silicatization and silicification. The association of sulphides with late-stage processes in pegmatitelike material is evidence that they formed a phase of the pegmatitic or volatile-rich granite magma. This "pegmatitic" material is considered to have been responsible for the contact metamorphism and migmatization as well.

INTRODUCTION

The high-temperature complex base metal sulphide deposit on Calumet Island, Quebec, is of special interest because its genesis is so closely related to "pegmatitization" and the general metamorphic history of the country rock. The ores, which carry zinc, lead, copper, gold, and silver, form roughly lenticular replacement bodies in a complex series of metamorphosed Grenville sediments. The mineralization is most abundantly developed in carbonate lenses and in contact metamorphic silicates derived from the carbonate. It is intimately connected in time and locus of introduction with silicatization of carbonate, silicification, and, probably, the introduction of alkaline material. Concentration of the metallic constituents is believed to have been favored by the physical properties of the country rock, especially the impermeability of carbonate-rich rocks under the conditions of mineralization, and by the relative permeability of associated thinly bedded siliceous gneisses. Mineralization has taken place during the final stages of injection and granitization, believed to be related to large granite intrusions to the east.

Geologic studies have been made of the Calumet deposit by Uglow (1916, p. 5), by Goranson (1925, p. 105), reviewed by Alcock (1930, p. 121), and Osborne. Uglow classed the country rock as mica gneisses, sericitic quartzites, probably of sedimentary origin, and limestone. Goranson considered that the deposits were located in biotitized shear zones in amphib-

olite, most of which he held to be igneous. According to his view, the carbonate and lime silicate lenses in which some of the best ore is located formed inclusions in the amphibolite. As already indicated, most, if not all, of the country rock is believed by the writer to represent sediments metamorphosed and injected by granitic material. There is little or no evidence available to indicate that shear zones other than the regional foliation have exerted a controlling influence on the production of mineralization.

Most of the material on which the present study is based was collected in connection with detailed mapping in the vicinity of the mineralized area on a scale of 100 feet to the inch. Approximately 50,000 feet of diamond drilling has been done on the property, and all the subsurface data have been obtained from the drill cores. Information regarding the regional geology has been drawn from the regional map by R. W. Ells (1907), from Goranson's report, and from personal observation. F. F. Osborne studied the deposit and the immediate area for the Quebec Bureau of Mines at the same time as the writer and supplied much information on the regional and detailed relationships of the mineralization.

ACKNOWLEDGMENTS

The writer wishes to express his indebtedness to the late Prof. Philip Krieger of Columbia University for suggesting this study and for his helpful advice and criticism during field and laboratory work. He is also under obligation to Profs. P. F. Kerr and S. J. Shand of Columbia University and Prof. E. S. Moore of the University of Toronto for critical reading of the manuscript. Thanks are due to Prof. F. F. Osborne of McGill University for helpful advice in the field and on various other occasions.

To Mr. Heath Steele and Dr. T. G. Moore of the American Metal Co. the writer is grateful for their good offices in making the problem available, and for numerous favors in connection with field and laboratory studies. The field work was made possible by the cooperation and hospitality of Dr. Paul Armstrong, of Calumet Mines, Ltd., who was in charge of operations at the property. Dr. M. H. Haycock of the Canadian Bureau of Mines kindly supplied several polished sections of the ores which were used in the present study.

The laboratory work on which this paper is based was carried out in the Economic Geology Laboratories of Columbia University.

LOCATION, HISTORY, TOPOGRAPHY

The deposit is situated on lots 3 to 12, range IV, of Calumet Island, in the Ottawa River (Fig. 1). The island is 53 miles west of the city of

Ottawa and is within easy reach of the Canadian Pacific Railway whose nearest station is at Campbell's Bay, a few miles to the east.

The deposit was known as early as 1893 and has since passed through several hands. Various attempts have been made at underground explora-

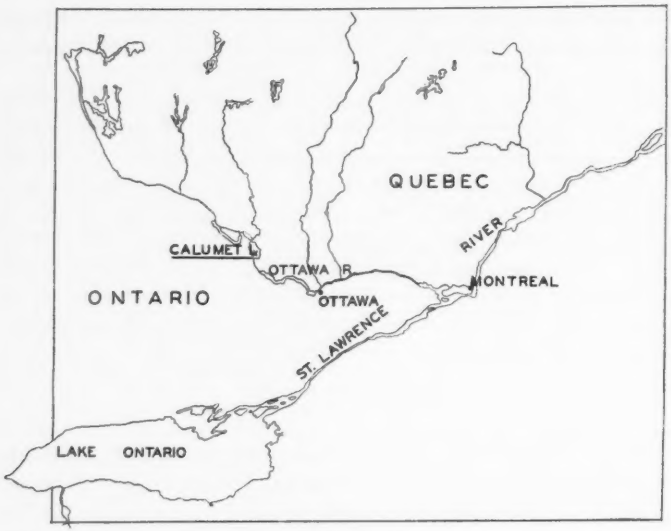


FIGURE 1.—*Location map*

tion, and about 1200 tons of ore have been shipped, mostly before 1900. In 1911 a concentrating mill with a capacity of 150 tons per day was erected but it was subsequently destroyed by fire. No really systematic attempt was made to develop ore until 1937, when an intensive diamond drilling campaign was instituted. Drilling continued during the summers of 1938 and 1939 and up to August 1939 had indicated an estimated 942,384 tons of ore, averaging 2.51 per cent lead, 8.16 per cent zinc, 0.036 oz. gold, and 5.76 oz. silver per ton (Osborne, 1939a, p. 1).

The western boundary of the accompanying geological map (Pl. 1) is formed by a branch of the Ottawa River. At this point limestone is exposed on the islands in the channel of the river and on the east bank. A steep north-south-trending scarp 100 feet high lies 100 to 300 feet east of the river. Glacial moraine obscures the bedrock from the cliff east to the mineralized area. East of the mineralized zone low outcrops of rock are abundant, interspersed with strips of swamp. Near the east boundary of the map area, the rocks are again partially hidden by deep drift.

GENERAL GEOLOGY AND STRUCTURE

GENERAL CONSIDERATIONS

A reconnaissance map by Ells (1907) indicates that the region in which the deposit is located is underlain by granitized gneisses and granite, carrying strips and patches of Grenville limestone of varying width. No detailed geological investigation of the general region has ever been made.

With the exception of the northwestern corner of the island, where Paleozoic sediments are found, the exposed rocks consist largely of a complex of amphibolites in various stages of granitization, with some areas of granite and limestone. The limestone forms an apparently continuous rim about the southern third of Calumet Island and dips uniformly to the east. This portion of the island is believed by Osborne (personal communication) to constitute the nose of a recumbent fold, possibly an anticline. If this be the case, the mineral deposit under discussion is located on the overturned limb of this structure.

In the immediate vicinity of the Calumet Island deposit, the rocks include limestone, "rusty" gneisses, amphibolites, and various migmatites. The contacts of these rock types follow a curving course, striking north-south or north-northwest near the northern boundary of the map and northwest-southeast near the southern boundary. In the eastern part of the area the contacts trend more uniformly north-northwest. They have a moderate inclination, roughly 30° E. Schistosity and bedding trend dominantly northwest-southeast and dip east at an angle averaging 30 to 40 degrees with local reverse dips.

A few hundred feet north of the map area the strike changes to nearly east-west. As already mentioned, an east-west component also appears in the strike to the south. Thus, the mineralized area is bounded by cross shears which have been developed over considerable distances, according to Ells' map. The localization of deformation along these shears, together with the arching of thick formations between them, both above and below the ore zone, probably protected the latter from the full impact of regional stresses. This may, in part, account for the varying degrees of metamorphism shown by the country rock of the deposit.

DISTRIBUTION AND STRUCTURE OF LITHOLOGIC UNITS

The lowest member exposed on the western margin of the map area is crystalline limestone which is at least 600 feet thick and strikes nearly due north-south. It is mainly coarse and massive but it contains sections of thin-bedded quartz-bearing dolomite and siliceous bands mapped as leucocratic gneiss, which probably represent original sandy beds.

The limestone is, for the most part, deformed by only minor warps and folds up to 100 to 200 feet in length. Just south of the junction

of Bowie Creek and the Ottawa River, however, the carbonate protrudes eastward about 250 feet. A little iron-stained siliceous gneiss overlies the nose of this sharp fold or "intrusion." During the development of this structure, the limestone underwent considerable flowage.

For the northern three quarters of its length the limestone belt is overlain by a thinly laminated, rather coarse-grained, siliceous, garnetiferous gneiss. The gneiss is injected to varying degrees by thin sheets of pegmatitic material. Foliation and *lit-par-lit* structure are locally considerably contorted. Amphibolite is developed along the contact of the gneiss with the limestone and is found as inclusions and bands within the gneiss itself.

To the north the band of gneiss strikes north-south, but approaching Bowie Creek it widens greatly, and the main belt trends northwest-southeast. Within the angle thus formed, the most massive facies is developed as light-cream, medium granular, oligoclase granite. Although for the most part conformable with the other units of the area, at this point the gneiss definitely crosscuts the amphibolite which lies south of it.

This amphibolite separates the limestone and garnet gneiss. To the south the amphibolite is divided by tongues of siliceous gneiss. It is made up largely of hornblende gneisses, much injected by granitic material and calcite, and is cut by pegmatite dikes. Locally, coarser-grained pyroxene-scapolite gneisses occur in this mass.

Roughly 1000 feet east of the main outcrop of garnet gneiss is a zone indicated on the map as "granitized rusty-weathering gneiss," characterized by predominantly rusty outcrops and clearly defined bedding. On the eastern border of the zone are found variable amounts of more massive, pargasite-bearing migmatites indicated on the map as "medium mafic granitized gneisses." Lenses of gneissic amphibolites occur in some abundance at various horizons. Bodies of pegmatite and massive quartz are numerous. Masses of limestone and silicated carbonate with which the most important mineralization is associated also outcrop within this complex group.

To the north, this mineralized zone is underlain by gneissic amphibolites. Near the north boundary of the area both have been deformed into a large drag foldlike crumple which has been exaggerated by flowage and injection. There seems to be some evidence of the "fold" in the distribution of formations to the southeast. In addition, numerous minor crumples are evident in well-exposed outcrops of these rocks. Carbonate masses in this zone appear to have undergone flowage, as indicated in some thin sections and exposures in the Lawn Pit.

A body of carbonate-bearing amphibolite, ranging from 100 to 300 feet in thickness, overlies this mineralized zone. The amphibolite consists

predominantly of green hornblende, quartz, and feldspar with disseminations, aggregates, and veins of calcite or lime silicates formed by metamorphism of the calcite. In many cases carbonate exceeds the amphibolitic material, and the rock becomes a breccia of amphibolite fragments in a matrix of calcite, instead of amphibolite veined with calcite. Owing to flowage of the enveloping carbonate the fragments are drawn out into spindlelike forms. The other extreme, amphibole schists with little or no carbonate, is also found in this zone.

The carbonate amphibolite zone is persistent and homogeneous. It shares to only a minor degree the injected and veined nature of gneisses which surround it. It is, however, moderately warped, and its bounding surfaces are extremely uneven.

To the north and east the "carbonate amphibolite" passes into more highly metamorphosed amphibolite and amphibolite gneiss. Where considerably granitized and injected, they are gneissic or even schistose and frequently garnetiferous. These amphibolites form rather massive outcrops which show, at times, a coarse foliation suggesting bedding.

Zones, streaks, and masses of granitized gneiss (migmatite) are complexly entangled with the amphibolites. The main bodies are situated just east of the carbonate amphibolite and are roughly triangular. They are separated from one another by a narrow zone of rather leucocratic hornblende gneiss. The most typical examples are light-colored, granite-like, garnetiferous, and highly siliceous, but more melanic varieties are also common, especially near the contact with amphibolites. The darker gneisses are classified on the map as "medium mafic granitized gneiss."

Although relict structures are sometimes in evidence, the foliation of siliceous gneisses of this type locally cuts across the bedding of less granitized gneisses to the south. This indicates that these migmatites were capable of intrusion and magmatic movement, at least locally.

South of the main zone of migmatites is a large well-exposed outcrop of less highly granitized iron-stained, well-bedded, badly contorted "medium-mafic" gneisses. They are injected pargasite-bearing rocks with minor intercalations of recrystallized amphibolite. Immediately east of these mixed rocks is a narrow double belt, locally interrupted by migmatite, which is composed of amphibolites and micaceous lenses of contact metamorphic origin, with some carbonate and siliceous gneiss. Some mineralization (the Belgian Pit and Longstreet Shaft) is associated with this zone.

To the east lie siliceous hornblende and biotite gneisses with lenses and bands of garnetiferous varieties and even-grained, granular, medium mafic amphibolites. East of Longstreet Creek massive coarse-grained amphib-

olites, of a relatively ungranitized nature, are abundant. Along the northern and eastern boundary of the property they have been strongly injected by granitic material to form medium mafic migmatites.

All the consolidated rocks and probably the mineralization are cut by a diabase dike approximately 100 feet wide at its widest part which trends roughly N. 80° W. to east-west and dips 70 to 80° N. There is some evidence that the dike is offset by a fault trending north-northwest near the east boundary of the area. A number of narrow trap dikes which are remarkably persistent and uniform in width were also intersected by drilling in the mineralized zone.

GENERAL STRUCTURAL FEATURES

The most characteristic feature of the structure is the prevalence of flowage. Evidence of flowage is found in the crystalline limestone, carbonate amphibolite, and various migmatites. Crumpling as well as some flowage affected thinly bedded siliceous biotite gneisses and some medium mafic migmatites. The most strongly granitized and siliceous migmatites show evidences of flowage and actual intrusion into less granitized rocks. The general east-west trend of the lineation indicates that the direction of flowage was generally the same in all the rock units. The lineation pitches 15° to 20° or occasionally even 45° E. The lineation is expressed in minute corrugations in lenses and drag-folded veins of quartz, drag fold axes, and the orientation of hornblende needles.

Jointing is abundant on a macroscopic and microscopic scale. Many of the joints are filled with granitic material. Thus, during at least the later stages of intrusion, some fracturing took place.

An important feature of the country rock of the deposit is the preservation of bedding structure in some of the quartzitic "rusty gneisses" and the persistence of less obvious structures probably of the same origin, in amphibolites and migmatites. The recognition of these structures is of special significance in the interpretation of the origin of the various rock types.

Some reference should also be made to the almost complete absence of evidence of deformation of any importance subsequent to metamorphism and mineralization. During metamorphism recrystallization was complete, for little or no trace of strain shadows in quartz, mortar structure, and other evidence of rupture is visible. The sulphides themselves show little structural disturbance apart from minor fracturing which took place during mineralization. This is significant, for rocks such as partially silicated carbonate and serpentine are easily deformed. Unfilled joints and the possible fault and associated fracture zones constitute exceptions to the absence of structural disturbances.

PETROLOGY

INTRODUCTION

The country rock of the Calumet Island ore deposit comprises four main groups: crystalline limestone and dolomite, amphibolites, biotitic "rusty-weathering" quartzites and paragneisses, and finally granitized gneisses or migmatites[1] of complex hybrid origin. The progress of metamorphism is best revealed in the amphibolites. Metamorphism began with the crystallization of hornblende from the original sediment with the separation of residual calcite. Contact metamorphism of the carbonate followed, caused and accompanied by injection. In the siliceous sedimentary gneisses, only thermal metamorphism and subsequent injection and granitization are recognizable. The progressive increase in intensity of metamorphism and injection thus represented indicates the progressively more intense action of intrusive granite.

The material added to the original sediments during the metamorphism includes silica in unusual quantities, alumina (in feldspars and possibly also in sillimanite and garnet), alkalies, and minor amounts of chlorine, phosphate, zirconium, fluorine, and water. Magnesia has been concentrated at least locally (as cordierite, biotite, anthophyllite, and pargasite), but there seems no necessity for assuming its introduction from an outside igneous source. Dolomites are abundant in the associated limestone masses and could easily supply magnesia to magmatic solutions coming in contact with them.

The most characteristic substances added are thus the constituents of a siliceous pegmatite. The ability of the injected material to penetrate along schistosity and bedding planes indicates that it must have been highly fluid, a property to be expected in late-stage pegmatitic material. The fact that many pegmatitic bodies in migmatites and amphibolites have gradational boundaries is significant in this connection.

The localization of the "pegmatitic" material during injection was dependent on several factors, including porosity, ability to fracture, and the abundance of shear surfaces such as bedding and schistosity. As a result of recrystallization and flowage during intense metamorphism, pure carbonate rock showed little susceptibility to injection, since it would neither fracture nor maintain pore spaces. Thus, in the area studied the limestone was injected only where it was thinly bedded or in contact with migmatites. The carbonate-bearing amphibolite, for the same reason, has not been subjected to extensive granitization.

[1] The term migmatite is here used in accordance with the definition of Sederholm (1907, p. 110)—i. e., a gneiss which has resulted from the mixture of intrusive and sedimentary or volcanic material. Its application is, however, largely confined to the more leucocratic siliceous mixed rocks, usually carrying garnet.

On the other hand, the "rusty-weathering" gneiss zone apparently possessed all the qualities enumerated and consequently has been the site of considerable granitization, silicification, and mineralization. Similarly, the production of migmatites appears to be related to the presence of bedding and shear planes. Where actual magmatic movement of the hybrid rock has taken place, the traces of such control may be destroyed.

The rocks of the area reflect considerable differences in the temperature and activity of the agents of metamorphism. As will be pointed out, the characteristic association of the carbonate amphibolite zone—calcite and hornblende—is stable only under intermediate conditions of metamorphism. On the other hand some of the siliceous migmatites, with labradorite and garnet, are to be classed with the higher-grade zones of metamorphism of Barrow and Tilley (Harker, 1939, p. 218). Sillimanite is also said to indicate the existence of high temperatures. This variability of grade of metamorphism within the same body of rock is the logical consequence of injection metamorphism, in which various zones are more susceptible to injection than others and thus are exposed to higher temperatures and more active agents. In conclusion, the metamorphic rocks to be described may be aptly termed a "polymetamorphic" (Wegmann and Kranck, 1931, p. 93) association. Variations in the grade of metamorphism are complicated by differences in the original sediments. The complexity has been increased by the development of reaction zones between carbonate-bearing and siliceous sediments, by varying degrees of injection by magmatic material, by differences in the degree of assimilation shown by the latter, and by the contact metamorphism of limestone interbeds under the action of the injected material.

CARBONATES

The limestones of the main western belt are clearly sedimentary and include both calcite and dolomite. Quartzitic interbeds, which have reaction borders of diopside and tremolite, may be assumed to represent original sandy horizons. Thin-bedded, siliceous dolomites form intermediate types between the quartzites and massive limestones. Silicates in disseminations, pockets, and bands are common.

The carbonate lenses in the mineralized zone of "rusty-weathering" gneisses, as exposed in various workings, are very irregular in their boundaries, and apparently unconformable. This is a consequence partly of the nature of the contact between silicated and unsilicated carbonate and partly a result of actual injection of the carbonate into the country rock during metamorphism. While definite generalizations are not possible until underground work has contributed more exact information, the general persistence of carbonate and lime silicate zones roughly parallel to

the structure of the associated rocks favors the interpretation of them as intercalated carbonate beds. The injection and replacement phenomena shown by the calcite would, according to this view, be considered merely a result of the plasticity and solubility of the mineral in question.

Minerals recognized in the contact rocks resulting from the silicatization of these carbonate lenses include anthophyllite, tremolite, actinolite, cummingtonite, pargasite, plagioclase (of varying composition), scapolite, phlogopite and biotite, and anisotropic garnet and spinel. Alteration products include serpentine, sericite, chlorite, epidote, clinozoisite, and wilsonite (Osborne, personal communication). Quartz has been introduced, locally in considerable abundance.

The limestone lenses have been silicated to a much greater degree than has the main marble belt to the west. This is believed to be due to their relatively small size and to the fact that they are more or less entirely surrounded by gneisses and migmatites which are the product of extensive injection and pegmatitization. Thus, a large surface, relative to their volume, was in contact with the injected material.

Where contact metamorphism of carbonate has taken place, it has often been followed or accompanied by injection. Various injected, coarse, biotitic rocks in the mineralized zone have probably been formed in this way. A very characteristic type, at the contact of the garnet gneiss and limestone, carries pyroxene, alkaline amphibole, biotite, scapolite, sericitized plagioclase, quartz, sphene, and carbonate, with varying amounts of microcline or orthoclase and often tourmaline. Goranson (1925, p. 111) has described rocks of this type as one of a series of zones of contact metamorphism of limestone. Similar mineral associations are common in the amphibolite mass which separates the limestone and garnet gneiss in the southwest corner of the area.

AMPHIBOLITE

The least metamorphosed representatives of this group are found in the well-defined zone of carbonate amphibolite. Characteristically, they are composed of amphibole and calcite, with minor quartz and feldspar and sometimes considerable biotite. As pointed out by Barth (1936, p. 791), this association represents an intermediate grade of metamorphism. With increased temperature and injection, this association became unstable, and diopside, pargasite, scapolite, quartz, and feldspar replaced the veinlets and aggregates of carbonate in the amphibolite more or less completely (Pl. 2, fig. 1). The presence of scapolite, quartz, and feldspar is proof that increased temperature alone did not produce this change but that metasomatism and injection co-operated in developing the contact silicates mentioned.

In a few specimens, discontinuous chains of hornblende form rims about calcite aggregates (Pl. 2, fig. 2) or contact metamorphic minerals replacing the calcite (Pl. 2, fig. 3). This implies that the amphibole is later than the carbonate. The "injection" of the amphibolitic material by the carbonate is thought to be due to flowage following the development of the hornblende. According to this view the calcite would be a residual from the formation of amphibole in the metamorphism of an impure dolomitic sediment. The formation of the hornblende may have been due to metasomatism of an intermediate grade or to regional metamorphism. The former has certainly been the case in the production of alkaline amphibole and actinolitic hornblende at the contact of garnet gneiss and limestone near the southwest corner of the map.

With a decreasing percentage of calcite the carbonate amphibolite passes into amphibole schists and hornblende rocks carrying little else but interstitial quartz and feldspar. These may result either from the recrystallization of a sediment whose composition was such that residual calcite was negligible or from the leaching of residual calcite during metamorphism by solutions traversing the rock.

With a greater percentage of plagioclase (andesine-labradorite) and quartz the above grade into even-grained "pepper-and-salt" amphibolites and amphibolite gneisses. Biotite, although frequently insignificant, is abundant in some specimens (Pl. 2, fig. 4). Diopside is occasionally present, and in the more granitized varieties garnet may be important. In contrast with that in the carbonate-bearing types, the hornblende in these amphibolites is darker and, when alkali feldspars are associated with it, distinctly bluish, in extreme cases approaching hastingsite.

Hornblende and biotite are frequently replaced by thin fingers of quartz and to a less extent by feldspar with the development of a pseudopoikilitic structure. Reference is made by Buddington (1939, p. 177) to similar structures in the Adirondacks, which he interprets in the same way. In advance stages of replacement by feldspar and quartz the melanic minerals may be tattered, ragged, and even sievelike. Ferromagnesian material thus replaced may recrystallize as small prisms sprinkled through the replacing minerals. Quartz is interstitial to plagioclase, sometimes replacing it.

Pargasite (or a related optically positive amphibole) is common in small amounts and is believed to have been formed by the bleaching of green hornblende. A comparable alteration of hornblende to cummingtonite is recorded by Eskola (1915, p. 225, 257).

With respect to the origin of the amphibolites in the vicinity of the deposit Goranson (1925, p. 116) states: "Locally, the original minerals

and texture of the amphibolite are preserved and it is known to be an altered hypersthene gabbro or diorite." It would thus seem that he regards the amphibolite as an altered basic intrusive. Elsewhere (p. 109) he states that in many cases evidence of origin of the amphibolite is obscure and that some are "hybrid rocks resulting from limestone."

From the evidence at hand the writer is disposed to regard most, if not all, of the amphibolites and hornblende gneisses exposed in the area as of primarily sedimentary origin. This interpretation is supported by the local occurrence of bedding in the amphibolites, the intercalation of amphibolite with other lithologic types, the absence of iron oxides common in amphibolites derived from igneous rocks, and by textures and structures in the carbonate amphibolites suggesting the metamorphic origin of the rock. The possibility of this origin of amphibolites is demonstrated by the endomorphic and exomorphic development of amphibolite at the contact of garnet gneiss and limestone and the replacement of a siliceous dolomitic zone by amphibolites near the southwest corner of the property. The metamorphism of carbonate-bearing sediments to form amphibolites in the Grenville region of Canada and in the Adirondacks is well known as the result of work by Adams and Barlow (1910, p. 99, 164) and Buddington (1939, p. 11).

The carbonate amphibolite and the zone of biotite gneiss to the west are separated by a discontinuous irregular belt of pargasite- and anthophyllite-bearing migmatites. These migmatites are fine- to coarse-grained and granoblastic or gneissic in texture. They usually weather to a light color and carry abundant plagioclase ranging from albite-oligoclase to labradorite. Garnet is common in some types, and hornblende and biotite may or may not be present. These rocks grade westward into pargasite-bearing biotite gneisses and eastward into unaltered amphibolite. They appear to constitute a reaction zone characterized by the addition of magnesia to the biotite gneisses and the withdrawal of iron from the amphibolite.

"RUSTY-WEATHERING" GNEISSES

The "rusty-weathering" gneisses are typically thinly bedded, iron-stained, much injected, and silicified. The less granitized are relatively fine-grained and apparently quartzitic. Phlogopite and scattered iron sulphides with some galena and sphalerite are universal. Sillimanite (Pl. 2, fig. 5) is often found in fine clumps of fibers or somewhat coarse spindles as replacements of biotite. More or less poikiloblastic, altered cordierite is present in some of these rocks. The most strongly recrystallized of the sillimanite-bearing rocks are pegmatitic microcline-mica gneisses. The finer-grained gneisses frequently carry abundant small

flakes of graphite. A labradorite-phlogopite hornfels (Pl. 2, fig. 6) is coarser-grained but characteristic of the less injected varieties.

Injection and recrystallization has affected all examples studied to varying degrees and has involved the introduction of quartz and, in the final stages, alkali feldspar. The latter replaces original silica, sometimes completely, and is interstitial to coarse recrystallized, veinlike quartz. In the fine-grained varieties with abundant alkalic feldspar, the sillimanite is often sheathed with sericite or muscovite (Pl. 2, fig. 5). The sillimanite itself is developed by recrystallization of aluminous material (original clay or derived biotite) under conditions accompanying the injection of pegmatites (Barth, 1936, p. 799).

Gneisses resembling the various sillimanite- and cordierite-bearing types described above are known as country rock in the base metal deposits of Åmmeberg, Sweden (Johansson, 1910). Broken Hill, New South Wales (Andrews, 1922), Bodenmais, Bavaria (Weinschenk, 1900), and Montauban, Quebec (Osborne, 1939b, p. 712). Osborne (personal communication) reports that the cordierite-anthophyllite association, typical of many Fennoscandian deposits (Orijärvi and Falun), also occurs at Calumet although it does not appear to be abundant. Many workers (Eskola, 1915, p. 262; Magnusson, 1936, p. 332; Osborne, 1939b, p. 726) consider these minerals to be due to "magnesia metasomatism." In the Calumet area, the abundance of dolomitic carbonate precludes the necessity of assuming the introduction of magnesia from magmatic sources.

The zone of "rusty-weathering" gneiss has been ascribed by Goranson (1925, p. 119) to the silicification and biotitization of amphibolite. As already noted, however, these rocks have a well-developed bedding structure, which is not the same as that shown by the amphibolite. A hydrothermal origin for these types also seems unlikely from a mineralogical point of view. The most typical alteration product of the amphibolite near the zone of "rusty-weathering" gneiss is pargasite and not biotite, and sillimanite and cordierite are not normal alteration products of basic rocks. Although graphite is not always of undoubted sedimentary derivation, its exclusive occurrence in these well-bedded siliceous rocks also supports the sedimentary origin of the rusty gneisses. The fine-grained nature of some of the quartz in the less recrystallized varieties is also unfavorable to a theory involving hydrothermal replacement by quartz and biotite in the deep vein zone.

The writer therefore believes that these rocks have been derived from thinly bedded, siliceous sediments which probably contained some carbonate and clay. The mineral assemblage and textures of this group are the

product of intimate injection, accompanied by varying degrees of metasomatism and the formation of contact metamorphic minerals.

The extensive injection and granitization of this zone, with the widespread occurrence of sulphide mineralization, are believed to be evidence of its relative permeability to granitic and hydrothermal solutions. This may have been due, in part, to its arenaceous and hence porous character, in part to its relative rigidity so that it fractured readily, and, in part, to the abundance of bedding planes and shear surfaces.

GRANITIZED GNEISSES OR MIGMATITES

The granitized gneisses, as a group, are extremely variable in mineral composition, a characterisitic resulting from variations in the amount and type of injection and the kind of rock which underwent migmatization. Quartz is usually the most important mineral except in a few cases where microcline or albite-oligoclase more or less completely usurps its place. Plagioclase, ranging from andesine to labradorite, is next in importance, and hornblende, biotite, and garnet are present in variable amounts. Pyroxene, epidote-clinozoisite, and pargasite were found in some varieties, and in one or two cases hypersthene, enstatite, or anthophyllite were observed. The variations in color and composition shown by these rocks are reminiscent of the endomorphic effects of the Pyrenees granite, described by Lacroix (1898-1899, p. 10, 40).

These migmatites are usually foliated and sometimes show *lit-par-lit* banding. They frequently give evidence of the introduction of quartz in considerable amounts, replacing plagioclase, biotite, and hornblende. Slender stringers, lenses, patches, and *lit-par-lit* injections of quartz visible in the outcrops of these rocks corroborate the microscopic evidence of the late introduction of silica.

Unusual textures are shown by many of the migmatites. Poikiloblastic labradorite swarming with rounded quartz grains is common in some very siliceous gneiss (Pl. 3, fig. 1). Garnet very frequently has a skeleton or poikiloblastic form, and anthophyllite, when present, has a more or less well-developed sieve structure.

These granitelike rocks are not considered to be true igneous rocks, but the products of injection and assimilation. Consideration of the following facts makes this apparent: (1) *lit-par-lit* structures, indicating planar structures preceding and controlling injection; (2) the presence of relics of bedding, especially in the "medium mafic" migmatites, and to a minor extent in the siliceous types; (3) the abundance of garnet; (4) the generally conformable nature of the migmatites, intercalated with amphibolites and carbonate lenses; (5) the presence of intermediate types,

indicating relations of the migmatites to "rusty-weathering" gneiss, amphibolites, and contact metamorphic rocks; (6) the changes in the properties of the minerals roughly paralleling the progress of injection. This includes the general, though not universal, increase in color of hornblende, the increase in index and abundance of pleochroic haloes shown by biotite, and the occurrence of garnet in the more highly granitized types; (7) the occurrence of basic plagioclase (labradorite) in very siliceous types has not been observed in a normal igneous rock. In spite of this, the migmatites show local intrusive relations. This is additional evidence that injection and not replacement alone took part in the formation of these gneisses.

The original sediment from which the migmatites were derived is not always clear. Gneisses which carry lime silicates with abundant potash feldspar are obviously an injection or assimilation product of limestone. Very well-banded migmatites are probably granitized and recrystallized "rusty-weathering" gneisses. The more common types, in which hornblende and biotite both occur, with garnet, cannot be so readily classified. The mere presence of either the amphibole or mica is not by itself significant since recrystallization may result in the formation of one instead of the other, depending on the conditions, such as the abundance of volatiles.

Migmatization thus appears to have been occasioned by the injection of dilute pegmatitic solutions along shear planes and bedding planes. According to this view this tenuous "magma" reacted with and replaced the original constituents. Where these reactions did not get to completion, or a second stage of injection set in, *lit-par-lit* gneisses resulted.

This is similar in its main features to the process of granitization as outlined by Goranson (1925, p. 114). Wegmann and Kranck (1931, p. 96), and Fersmann (1931, p. 72) also consider migmatites to be the result of the action of pegmatitic solutions on older rocks.

A very common feature of the amphibolites and migmatites of the area is the occurrence of numerous pegmatitic zones, usually rather narrow, which carry the same minerals as the surrounding gneisses. They differ in the coarseness of grain, in the abundance of quartz, and in the frequent occurrence of coarse pink garnet and clove brown anthophyllite. Plagioclase is usually oligoclase or andesine, in well-twinned large interlocking grains. Frequently a little sulphide and various alteration products, such as sericite, serpentine, epidote, and chlorite, are present. A common feature is the recrystallization of the melanic constituents of the country rock of the pegmatite, along the borders of the latter, to form coarse grains of biotite, amphibole, or pyroxene. The same feature has been described by Goranson. Unusual microcline pegmatites of similar origin carrying

sillimanite and garnet or altered cordierite and tourmaline have also been recognized.

Microline-perthite pegmatites which, according to Goranson (1925, p. 110), are relatively abundant in other parts of the island are found as bodies with gradational boundaries and, occasionally, crosscutting dikes in the amphibolites and migmatites. Graphic granite was noted in some drill sections.

MINERALIZATION

THE ORE ZONE

The sulphide bodies of the Calumet deposit are simple in structure but rather complex in their mineralogy and lithologic association. The largest body (the Bowie) closely underlies the footwall of the carbonate amphibolite. It consists of a series of lenses of carbonate and contact silicates, which are usually intercalated with injected material, consisting of siliceous biotite gneiss, pegmatite, and clear massive quartz.

The sulphide bodies of the Russell and McDonald area are located well within the zone of "rusty-weathering" gneiss and are of the same general nature as the Bowie although much smaller. The limestone and lime silicate lenses in which they are developed occur at several different horizons and do not constitute a single, definite zone. On the other hand, low-grade, disseminated sulphide mineralization is quite widely distributed in the typical siliceous "rusty-weathering" gneisses.

The mineralization in the Longstreet zones also occupies a contact-metamorphosed carbonate horizon containing intercalated migmatites. Mineralization in limestone near the southwest corner of the map comprises a few thin streaks of sulphide in a dolomitic bed.

The three principal ore bodies discovered to date all dip east, conformable to the general structure of the country rock (Fig. 3). They are lenticular in cross section and roughly triangular in plan (Fig. 2). The long axis of the Bowie body trends northeast, whereas the Russell and McDonald bodies trend more nearly east-west. Thus, they tend to converge toward the east, presumably in the direction from which the mineralizing solutions came.

The structure of the bodies in question is complicated by minor cross folding. This is especially well shown by sections of the Bowie taken perpendicular to its major axis (Fig. 2). Distinct "rolls" are apparent also in the plane of the major axis. This warping is probably pre-ore. The thickest mineralized sections of the Bowie body are associated with convex upward or "anticlinal" parts of the rolls, in both longitudinal and transverse sections. Toward its eastern extremity the dip of the Bowie

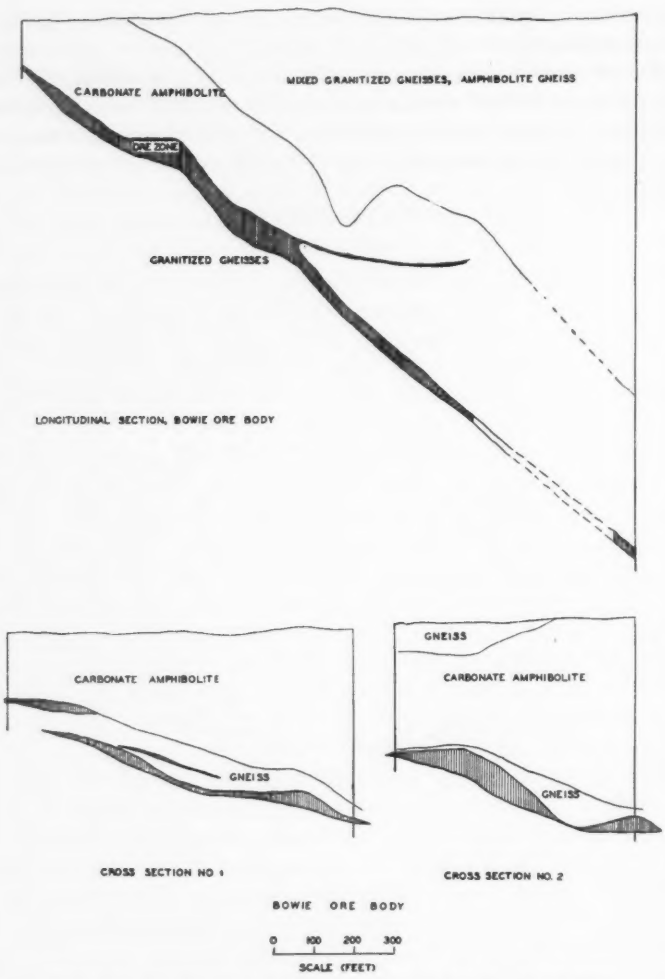


FIGURE 2.—*Longitudinal and cross sections of the Bowie ore body*
Based, in part, on drill logs by F. F. Osborne.

ore body varies less and is somewhat steeper. The mineralization is considerably thinner here, where changes in the dip are absent.

The boundaries of the ore zones roughly parallel the east-west lineation or flowage directions of the country rock. This is to be expected since the

carbonate lenses which were later mineralized must have been elongated parallel to this direction during deformation. Thus, the form of the sulphide bodies has been partially determined by mechanical flowage during metamorphism, by the intensity and distribution of cross folds and

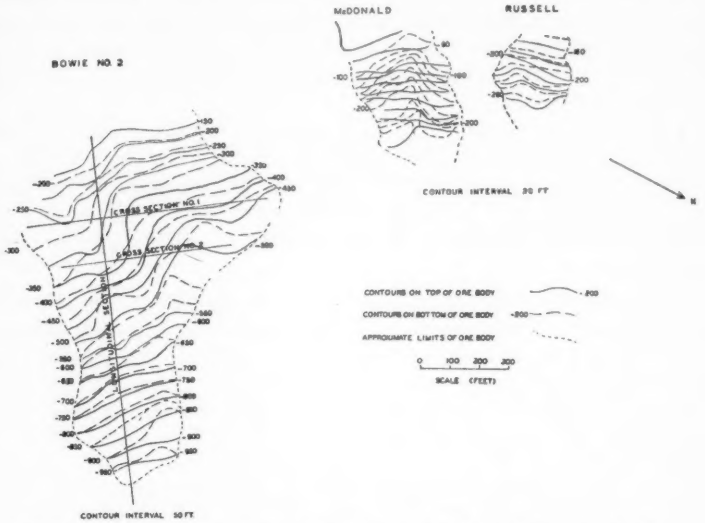


FIGURE 3.—*Contours on top and bottom of the three known ore bodies, measured from an arbitrary datum*

Figures to right of each body give depth of contour of top; figures to left give depth of contour of bottom.

"rolls," and, to a certain extent, by the usual boundary irregularities of metasomatic replacement.

The largest concentration of sulphides (in the Bowie body) may be partially attributed to the overlying "blanket" of carbonate amphibolite. The amphibolite is considered to have been relatively impermeable to granitizing and mineralizing solutions and thus formed a traplike structure, facilitating the concentration of metallic constituents and migmatites directly below it. This, of course, assumes that the incoming magmatic material had a tendency to rise vertically as well as to move up the dip. The fact that the sulphide body tends to be thicker where "rolls" result in flattened or reverse dips, forming low noses and anticlines, affords some confirmation of this assumption. It is alternatively possible that the concentration of fractures or foliation planes in the crest of the anticlinal "rolls" made these points structurally more favorable for ore deposition.

It seems unnecessary to assume any channelway for the introduction of mineralization and magmatic material other than the "rusty-weathering" gneiss. No evidence has been so far observed of any important shear zones, fractures, or joint systems which might have served such a purpose. On the other hand, the abundance of silicification, migmatization, injection of pegmatites, silication of limestone lenses, and widespread sulphide mineralization within the rusty gneiss has already been mentioned as evidence that it has served as a major conduit and a receptacle for introduced magmatic material.

A similar function may have been performed by the mineralized siliceous cordierite-anthophyllite gneiss and overlying amphibolite at Orijärvi, Finland (Eskola, 1915, p. 234), the median gneiss in the limestone at Balmat and Edwards (Brown, 1936, p. 238, 245), and the pyritiferous rusty gneisses in St. Lawrence County, New York (Smyth, 1912, p. 144).

Some reference should also be made to the distribution of ores with respect to the injected material. Granitic and pegmatitic material is so abundant that no critical correlation can be made. The most abundant mineralization, however, overlies the principal belt of garnetiferous migmatite, and the minor mineralization of the Longstreet area overlies a small, but clearly defined body of massive, siliceous, garnetiferous gneiss. A spatial relationship to the migmatite is shown by the mineralization in the limestone along the river which lies near the southern tip of a spur of the main garnet gneiss horizon.

PARAGENESIS OF THE ORES

General statement.—The principal sulphide minerals, in order of their paragenesis, are: pyrite, sphalerite, pyrrhotite, galena, tetrahedrite, chalcopyrite, and marcasite. Pyrite was undoubtedly the earliest to crystallize; the others are not so clearly separated from one another in the sequence but appear to have overlapped considerably. "Emulsion" structures of chalcopyrite and pyrrhotite in sphalerite and complex mixed grains of galena, sphalerite, and pyrrhotite indicate at least partial contemporaneity of all the sulphides except pyrite.

The late crystallization of the chalcopyrite and galena is probably responsible for the marked tendency which they show to make their way further into the gangue than other sulphides (Pl. 3, fig. 2). Marcasite appears to be the latest of the metallic minerals since it cuts and replaces all other sulphides.

Native gold (Krieger, personal communication) has been reported in ores which carry high precious metal values. Arsenopyrite has been observed and from its relationships probably is to be associated with pyrite

in time of formation. Bournonite was observed in the galena in one polished section. A little molybdenite was noted in fine plates scattered along an altered joint plane in injected biotite gneiss. Its relations to the other sulphides are unknown.

Sulphides are not uniformly distributed through the mineralized areas. They may be massive over short sections, although more often they occur as irregularly distributed blebs and pockets and as streaks, clots, and grains scattered through gneiss, quartz, and lime silicate minerals in carbonate.

Pyrite.—Pyrite occurs as euhedral, subhedral, and anhedral grains. Frequently it is surrounded by pyrrhotite, sphalerite, galena, or marcasite (Pl. 3, fig. 3), and in some specimens it has undergone considerable alteration to marcasite. Although it is fairly persistent it is never the dominant sulphide.

Arsenopyrite.— In a few specimens arsenopyrite was observed in occasional subhedral small grains, either intergrown with pyrite or enclosed in other later (?) sulphides.

Sphalerite.—Sphalerite is usually dark brown and ferriferous. Some however, occurs as white irregular grains replacing gangue around the rim of larger brown sphalerite masses. A similar light-colored sphalerite, which is later than the more abundant ferriferous material, has been described by Smyth (1917, p. 26) from the Edwards Mine, New York. Where the sphalerite has been replaced by or is associated with marcasite it has been bleached white where the two come in contact with one another. In some massive ores, colorless sphalerite is found intergrown with quartz.

When massive, the sphalerite is generally coarse-grained. A peculiar pseudoporphyritic type was found in the Bowie Pit. The "phenocrysts" are large areas of pure sphalerite, whereas the matrix is a fine-grained mixture of sphalerite with galena and pyrrhotite (Pl. 3, fig. 4). When associated with considerable galena, the sphalerite is sometimes intimately intergrown with it (Pl. 3, fig. 2).

Sphalerite frequently carries numerous small inclusions of chalcopyrite and pyrrhotite which resemble exsolution structures. Although chalcopyrite has been frequently described in this relationship, pyrrhotite (Pl. 3, figs. 5, 6) is much less well known. Schwartz (1937, p. 31) has reported a similar "emulsion structure" of pyrrhotite in sphalerite from the Potosi Claim, Chihuahua, Mexico. Veinlets of pyrrhotite and chalcopyrite also occur cutting the sphalerite, indicating that they are at least partly younger than their host.

Pyrrhotite.—Pyrrhotite is a ubiquitous and characteristic sulphide usually occuring as coarse, irregular blebs. Its relationship to sphalerite and pyrite have already been mentioned. When massive, it is often intergrown with sphalerite with mutual boundaries.

In most of the material studied, the pyrrhotite is partially replaced by marcasite, forming lattice, grid, or concretionary structures (Pl. 4, figs. 1, 2). The replacement is accompanied by the development of cavities at the contact of the two minerals. The openings are filled by later gangue, usually carbonate, and are sometimes bounded by euhedral marcasite (Pl. 4, fig. 3). In some cases, in the initial stages of replacement, the pyrrhotite assumes a peculiar pale grayish-white color (in reflected light) and exhibits a gray-black anisotropism. This alteration borders interrupted, anastomosing cracks due to shrinkage accompanying the alteration.

Galena.—Galena is generally associated with other sulphides, particularly sphalerite. Like most of the minerals already described, it may form coarse masses intergrown with sphalerite, often with mutual boundaries, or it may be finely disseminated through the gangue. It sometimes occurs as delicate rims surrounding grains of silicates and carbonates, with slender fingerlike projections branching out along cleavages, fractures, and grain boundaries (Pl. 3, fig. 2). Small quantities of tetrahedrite are consistently associated with galena. As already mentioned, in one specimen bournonite was noted, distributed through the galena.

The silver content of the ore is found in the galena and tetrahedrite. In the galena it is present as fine-grained dots of argentite or polybasite (Krieger, personal communication) visible under high power after etching.

Chalcopyrite.—Chalcopyrite is rarely as abundant as the minerals already discussed, and it is not as consistently represented in all of the specimens examined. It is usually found as fine-textured, disseminated grains and thin veinlets associated with other sulphides, chiefly sphalerite. In some of the more massive ore it may be quite coarsely crystalline. Reference has already been made to occurrence of chalcopyrite as an "emulsion structure" in sphalerite. One instance of the same relationship to tetrahedrite was observed.

Marcasite.—Marcasite shows a variety of habits (Pl. 4, figs. 1, 2, 3). Although most often associated with pyrrhotite and pyrite, it also replaces other sulphides such as chalcopyrite, sphalerite, and galena to a minor degree. In the case of the Calumet ores, the fact that all the ore minerals, including marcasite, are cut by fractures which carry carbonate, serpentine, and fine-grained quartz, and the absence of any evidence of leaching, oxidation, or enrichment favor a hypogene origin for the mineral. Also,

the development of marcasite in pyrrhotite which is entirely enveloped in massive sulphides (Pl. 3, fig. 4) can scarcely be supergene.

Marcasite is reported from similar complex ores at the Bodenmais Mine, Bavaria, (Weinschenk, 1900, p. 65) and from a deposit in the Eastern Alps described by Friedrich (1932, p. 497). At the latter deposit it is considered to be supergene.

RELATIONS OF SULPHIDES AND GANGUE

The sulphide minerals replace most of the associated gangue minerals, including amphiboles, pyroxene, phlogopite and biotite, sillimanite, carbonate, quartz, feldspar, and sericite. They are developed in disseminations in silicate rock, in mixed silicate and carbonate rock, and in siliceous gneisses and pegmatites. Some specimens of massive ore contain inclusions of unreplaced gangue which consists largely of calcite, and it is quite possible that, in such cases, the principal mineral replaced has been relatively pure carbonate. The ores thus do not show so consistent a localization in zones of mixed silicates and carbonate as was found in other lead-zinc deposits by Brown (1940). In specimens of mixed carbonate and lime silicate the sulphides commonly appear to be more intimately associated with the silicates. The ore minerals show a consistent spatial association with the silicates, and both are more or less interstitial to the carbonate aggregates (Pl. 4, figs. 4, 5). Occasionally inclusions of sulphide are found within the silicate grains (Pl. 4, figs. 4, 5). These relations are thought to be not so much a measure of the relative ease of replacement as a result of the genetic relations between the sulphides and silicate minerals at Calumet. Where silicates, such as amphiboles, are the dominant gangue, the sulphides are located interstitially to them and in fractures and cleavages. Replacement of the silicates has undoubtedly taken place but does not appear to have been extensive.

The relationships shown by sphalerite and other sulphides to vein quartz are sometimes significant. The sulphides, in the cases under consideration, are interstitial to the quartz and fill fractures in it, suggesting that they have been introduced later. On the other hand, large grains of sulphide in this association are free from inclusions or unreplaced remnants of quartz. In one specimen of massive sulphides, quartz and sphalerite were noted in delicate intergrowth. Also, in some amphibole-rich lime silicate rocks, the most abundant sulphide mineralization accompanies small quartz lenses. These all suggest that the quartz and sulphides did not come in at different stages of mineralization but that they were introduced in the same solution. The apparent later age of the sulphides is believed to be due to their later solidification. The intergrowth mentioned between sphalerite and quartz indicates that there was some overlap in the range of

crystallization of the two. Similarly, massive sulphides occurring in streaks, veinlets, and blebs in mineralized pegmatite, although obviously emplaced after the crystallization of the pegmatite, are also thought to have been derived from the same igneous source and introduced as part of the pegmatitic liquid.

Sulphide minerals are sometimes associated with sericitization and albitization. In mineralized microcline pegmatite specimens from the Bowie Pit the sulphides are rimmed by albite and are not found in unaltered feldspar (Pl. 4, fig. 6). In some lime silicate ores also, sericitized patches of feldspar are dotted with tiny grains of galena.

Serpentine occurs as an alteration of pyroxene, phlogopite, and amphibole, as veinlets which cut the ore minerals, and as rims around marcasite. Thus, the serpentine was probably formed during the final phases of mineralization, in association with marcasite. Smyth (1917, p. 22) described a similar relationship at Edwards, New York.

PHYSICAL CONDITIONS OF MINERALIZATION

The temperature and pressure under which mineralization took place were probably high. The intimate association of lime silicate minerals, especially diopside and garnet, with the sulphides is indicative of a relatively high initial temperature of mineralization. Pyrrhotite, arsenopyrite, and ferriferous sphalerite are also characteristic of hypothermal, high-temperature conditions.

The flowage of the limestone and carbonate-rich rocks and the *lit-par-lit* injection in some of the gneisses indicate that high pressures must have existed during metamorphism and mineralization. These conditions probably prevented more than a local and temporary development of a vapor phase if, indeed, any developed at all.

NATURE OF MINERALIZING SOLUTIONS

Contact metamorphism of the limestone lenses in the ore zone was effected by the introduction of various substances to the limestone, which was probably dolomitic. The extent to which impurities may have supplied ingredients for the crystallization of the silicates is not known, but it is thought that their role was a minor one. Assuming that this is true, the substances introduced include: abundant silica, as quartz and as a constituent of feldspar and lime silicates; alumina, chiefly in feldspar, garnet, mica, spinel, and epidote or zoisite; soda and potash in mica, feldspar, and scapolite; phosphate in apatite; fluorine in apatite and perhaps in phlogopite and pargasite; titanium in sphene and rutile; zirconium in zircon; and abundant water in various silicates. During mineralization, sulphur and iron were the most abundant constituents introduced. Zinc, lead,

copper, antimony, arsenic, and minor silver and gold accompanied them.

Mention has already been made of the pegmatitic nature of the granitizing solutions. The inference may tentatively be drawn that contact metamorphism was largely the result of reaction between pegmatitelike solutions and carbonate lenses in the original sediments. Similar conclusions have been reached by Agar (1923, p. 119) in the Oxbow region of the West Adirondacks and by Emmons and Calkins (1913), who ascribe contact metamorphism in the Philipsburg, Montana, region to agents "related to pegmatites in composition" but more diffuse.

To sum up, the sulphides appear to have been closely associated with the solutions (probably pegmatitic in nature) which caused contact metamorphism with late-stage albitization of pegmatitic material and with vein quartz. Goranson (1925, p. 121) has emphasized the abundance of pegmatites in the Lawn and St. Anne areas, although he believes that the mineralization is later than the pegmatites. It is admitted that the sulphides generally crystallized later than the silicates and quartz of the pegmatites and contact metamorphic lenses and that they underwent transfer to varying degrees after the latter had been formed. It is felt, however, that the relationships just reviewed suggests that the solutions which caused contact metamorphism were pegmatitic in nature and that the sulphides were associated with the pegmatitic fluid. Owing to the mobility of the sulphides, the absence of a correlation between pegmatites and metallic mineralization is not critical, but, even if it is only local and occasional, the preservation of such an association is significant.

The preceding discussion implies that there is a closer relation between the sulphide ores and the pegmatites than is usually accepted (Lindgren, 1933, p. 708; Fenner, 1935, p. 60). If the migmatization and metamorphism are due to agents of a pegmatitic nature, then we are faced with the possibility that magmatic solutions which, under suitable conditions, crystallize as pegmatite dikes may, under other conditions, invest the country rock so intimately that their pegmatitic affiliations are concealed. Graton (1940, p. 228) has cited the inadequate amount of pegmatite in association with many ore deposits as evidence against the importance of pegmatites in sulphide mineralization. If the suggestions made above are true, this lack of pegmatites, in some cases at least, may be more apparent than real.

Most of the complex base-metal sulphide deposits which have been mentioned (Montauban, Edwards and Balmat, Orijärvi Ämmeberg, Eastern Alps, Broken Hill, and others) have pegmatites associated with them in some abundance. It is undoubtedly true that there is a genetic basis for this although it is not necessarily as close a relationship as seems to be true at Calumet.

The role of the limestone in the ore genesis is, likewise, significant. Goldschmidt (1911, p. 1) has described carbonate rocks as "absorption machines," a function which undoubtedly belonged to the lenses of ore-bearing carbonates in the Calumet deposit.

SOURCE OF PEGMATITE AND MINERALIZING FLUIDS

The metamorphism, injection of igneous material, and mineralization are evidence of the influence of intrusion and crystallization of a near-by igneous body. The presence of a small mass of oligoclase granite in the garnet gneiss zone, of larger zones of gneissic granite 1 or 2 miles east of the property, and a large area of gneissic batholithic granite 3½ miles east of the Calumet deposit may represent the granitic parent magma. Other intrusives, such as the mass of basic rock and derived amphibolite east of Bryson, Quebec, and the diabase dike adjacent to the mineralized zone show no evidence of being genetically related to the deposit.

GENERAL SUMMARY

The various processes of metamorphism, pyrometasomatism, metasomatism, and granitization are all considered here to be the effects of the injection and permeation of older sedimentary rocks by the late pegmatitic extracts of a granitic magma. The mineralizing solutions which produced the high-temperature complex sulphide deposit were related to and probably were associated with the pegmatitization, at least in its later stages. The mineralization thus forms an integral part of the metamorphic and igneous history of the area and constitutes the last important mineralogical change which the rocks of the area have undergone.

The succession of events may be tabulated thus:

1. Regional, or low-grade hydrothermal metamorphism
   (a) Formation of amphibole in impure dolomitic sediments, as first stage in the production of amphibolites, with the separation of residual calcite.
   (b) (Recrystallization and formation of mica in impure siliceous sediments as first stage of "rusty-weathering" gneisses. No examples preserved.)
2. Contact metamorphism
   (a) Contact silicates produced in carbonate amphibolite.
   (b) Formation of phlogopite-labradorite gneisses and fine-grained quartz-phlogopite gneiss, with recrystallization of excess carbonate to lime silicates.
   (c) Contact metamorphism of limestone lenses in "rusty-weathering" gneiss zone.

3. Injection and granitization (in part at least contemporaneous with and effecting stage 2, but subsequent to it at any one point).
   (a) Amphibolite gneisses and garnet migmatites.
   (b) Thinly banded biotite gneisses and garnet migmatites.
   (c) Coarse injected biotitic zones.
   The formation of pargasite from hornblende is thought to be associated with this stage. The production of sillimanite also probably took place during injection.
4. Sulphide mineralization associated with silicification, silicatization, injection, and albitization.
5. Formation of late serpentine, carbonate, and silica following minor fracturing of sulphides.

It should be emphasized that the subdivision into stages is an artificial simplification of processes which were often contemporaneous.

TYPE OF ORE DEPOSIT

The sulphides in siliceous, pegmatitic, and gneissic gangue are high-temperature metasomatic deposits, since they replace the silicate minerals at least to some degree. The sulphides associated with the limestone lenses show all the characteristics of pyrometasomatic or "contact metamorphic" deposits without a true igneous contact. However, it is obvious that they formed under the influence of the same agents which produced the first groups and therefore, as Graton (1933, p. 531) justly remarks, they may simply be regarded as representatives of the hypothermal zone in limestones.

WORKS TO WHICH REFERENCE IS MADE

Adams, F. D., and Barlow, A. E. (1910) *Geology of the Haliburton and Bancroft areas,* Geol. Survey Canada, Mem. 6, p. 99, 164.

Agar, W. M. (1923) *Contact metamorphism in the West Adirondacks,* Am. Philos. Soc., Pr., vol. 42, no. 3, p. 119.

Alcock, F. J. (1930) *Zinc and lead deposits of Canada,* Geol. Survey, Canada, Econ. Geol. Ser., no. 8, p. 121.

Andrews, E. C. (1922) *The geology of the Broken Hill District,* Geol. Survey, New South Wales, Mem. 8.

Barth, T. F. W. (1936) *Structural and petrological studies in Dutchess County, New York,* Geol. Soc. Am., Bull., vol. 47, p. 791.

Buddington, A. F. (1939) *Adirondack igneous rocks and their metamorphism,* Geol. Soc. Am., Mem. 7, p. 11, 177.

Brown, J. S. (1936) *Structure and primary mineralization of the zinc mine at Balmat, New York,* Econ. Geol., vol. 31, p. 238, 245.

——— (1940) *Factors of composition and porosity in lead-zinc replacements of metamorphosed limestone,* Am. Inst. Min. Metall. Eng., Tech. Paper 1194.

Ells, R. W. (1907) *Report on the geology and natural resources of the area included in the Northwest Quarter-Sheet No. 122 of the Ontario and Quebec Series comprising portions of the counties of Pontiac, Carleton, and Renfrew,* Geol. Survey Canada.

Emmons, W. H., and Calkins, F. C. (1913) *Geology and ore deposits of the Philipsburg Quadrangle, Montana,* U. S. Geol. Survey, Prof. Paper 78.

Eskola, P. (1915) *Petrology of the Orijärvi Region,* Comm. Geol. Finland, Bull. 40, p. 225, 234, 257, 262.

Fenner, C. N. (1935) *Pneumatolytic processes in the formation of minerals and ores,* in *Ore deposits of the Western States,* Am. Inst. Min. Metall. Eng., p. 60.

Fersmann, A. (1931) *Über die geochemisch-genetische Klassifikation der Granitpegmatite,* Mineralogische und Petrographische Mitteilungen, Band 41, p. 72.

Friedrich, O. (1932) *Eine alte pegmatitische Erzlagerstätte der Ostalpen,* Neues Jahrb. für Mineral. Geol. und Palaeont., Beil, Band 65, Abt. A, p. 497.

Goldschmidt, V. M. (1911) *Die Kontaktmetamorphose im Kristianiagebiet,* Vidensk. Skr., p. 1.

Goranson, R. W. (1925) *Calumet Island, Pontiac County, Quebec,* Geol. Survey Canada, Summ. Rept., pt. C, p. 105-121.

Graton, L. C. (1933) *The depth zones of ore deposition,* Econ. Geol., vol. 28, p. 531.

——— (1940) *Nature of the ore-forming fluid,* Econ. Geol., vol. 35, Supplement to no. 2, p. 228.

Harker, Alfred (1939) *Metamorphism,* London, p. 218.

Johansson, H. E. (1910) *The Ammeberg zinc ore field,* Intern. Geol. Cong., Sweden, 1910, Excursion C4.

Lacroix, A. (1898-99) *Le granite des pyrénées et ses phénomènes de contact,* Bull. Carte Geol. de France, vol. 10, no. 64, p. 10, 40.

Lindgren, Waldemar (1933) *Mineral deposits,* 4th ed., McGraw-Hill, New York and London, p. 708.

Magnusson, J. H. (1936) *The evolution of the Lower Archaean rocks of Sweden and their iron, manganese and sulphide ores,* Geol. Soc. London, Quart. Jour., vol. 92, p. 332.

Osborne, F. F. (1939a) *Preliminary report on southern part of Calumet Island and the adjacent mainland,* Quebec Dept. Mines, Prog. Rept. 140.

——— (1939b) *The Montauban mineralized zone, Quebec,* Econ. Geol., vol. 34, p. 712.

Schwartz, G. M. (1937) *Paragenesis of pyrrhotite,* Econ. Geol., vol. 32, p. 31.

Sederholm, J. J. (1907) *Granite and gneiss,* Comm. Geol. Finland, Bull. 23, p. 110.

Smyth, C. H., Jr. (1912) *On the genesis of the pyrite deposits of St. Lawrence County, New York,* N. Y. State Mus., Bull. 158, p. 144.

——— (1917) *Zinc ores of the Edwards District, St. Lawrence County, New York,* N. Y. State Mus., Bull. 201, p. 22-26.

Uglow, W. L. (1916) *Lead and zinc in Ontario and Eastern Canada,* Ontario Bur. Mines, vol. 25, pt. 2, p. 5.

Wegmann, C. E., and Kranck, E. H. (1931) *Beiträge zur Kenntnis der Svecofinniden in Finnland,* Comm. Geol. Finland, 15, no. 89, p. 93, 96.

Weinschenk, E. (1900) *Der Silberberg bei Bodenmais im Bayerischen Wald,* Zeits. für prakt. Geol., p. 65.

Department of Geology, University of Toronto, Toronto, Canada.
Manuscript received by the Secretary of the Society, July 8, 1940.

EXPLANATION OF PLATES 2-4

PLATE 2

PHOTOMICROGRAPHS OF AMPHIBOLITES AND GNEISSES

FIGURE 1.—CARBONATE AMPHIBOLITE

Portion of a veinlet of calcite (pale gray, showing twin lines) partially replaced by pyroxene (light gray, with strongly marked cleavage) and colorless quartz and feldspar. Amphibole occurs as darker gray grains at borders of the photograph and as one or two isolated grains in the center. Plain light. x35.

FIGURE 2.—CARBONATE AMPHIBOLITE

Chains of hornblende (black) surrounding granules of carbonate (gray), with quartz and feldspar. Plain light. x35.

FIGURE 3.—CARBONATE AMPHIBOLITE

Pyroxene (light gray, high relief, strong cleavage) and scapolite (white with faint cleavage) surrounded by a discontinuous rim of hornblende (dark gray). Quartz and feldspar (white) associated with hornblende. Biotite in upper right hand corner. Plain light. x35.

FIGURE 4.—BIOTITE AMPHIBOLITE

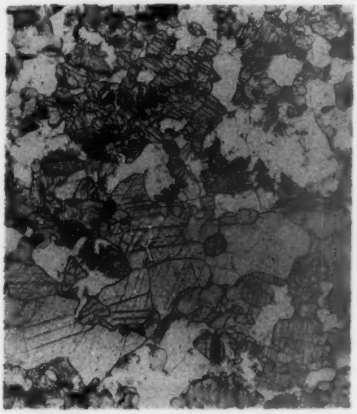
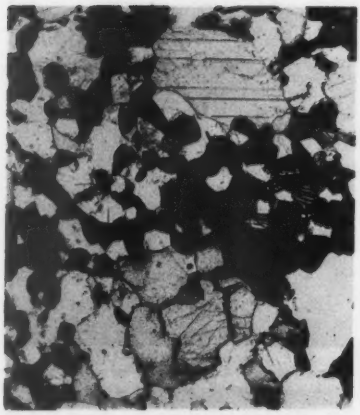
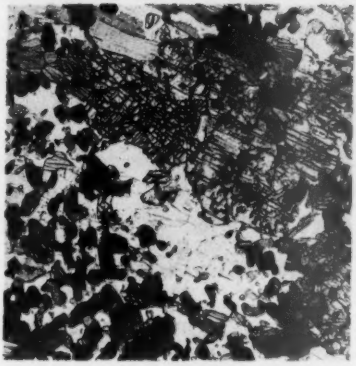
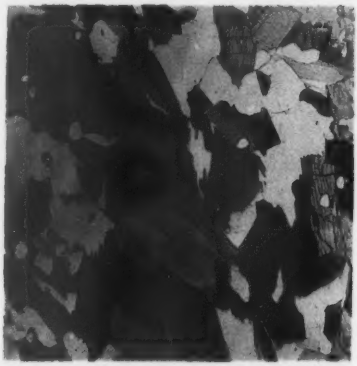
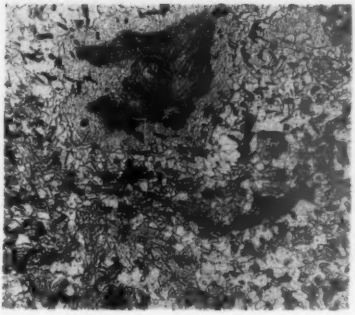
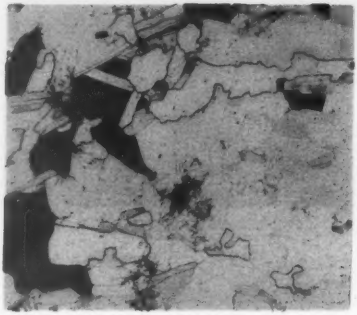
Hornblende (pale gray), biotite (medium gray to black), corroded by quartz and feldspar (white). Plain light. x35.

FIGURE 5.—FINE-GRAINED "RUSTY-WEATHERING" GNEISS (taken slightly out of focus)

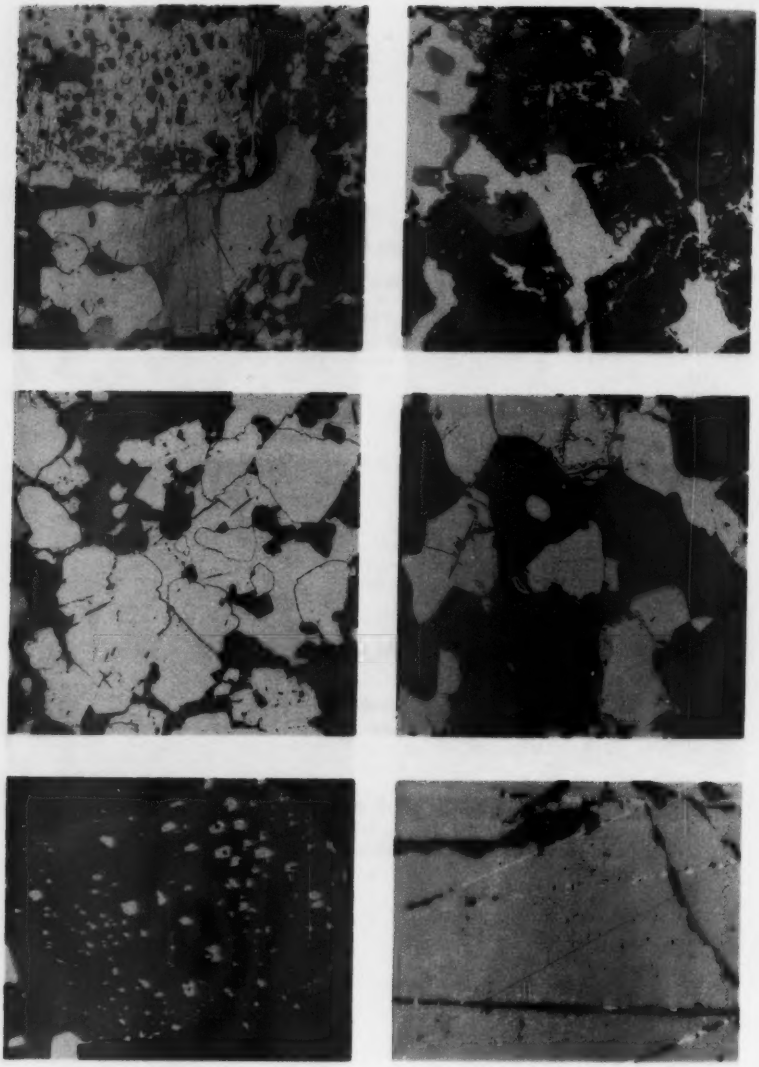
Sillimanite (fibrous, dark-gray mass) surrounded by granular-looking sericite. Quartz (white) with interstitial alkali feldspar (faintly gray mottled). Darker-gray, irregular material phlogopite. Sulphides black. Plain light. x35.

FIGURE 6.—"RUSTY-WEATHERING" GNEISS ZONE

Labradorite (pale gray) and biotite (medium gray with pleochroic haloes) cut by veinlike masses of quartz (outlined in ink). Sulphides (black) associated chiefly with the phlogopite. Plain light. x35.

PHOTOMICROGRAPHS OF AMPHIBOLITES AND GNEISSES



GNEISSES AND PARAGENESIS OF ORES

PLATE 3

GNEISS AND PARAGENESIS OF ORES

FIGURE 1.—SILICEOUS GARNETIFEROUS GNEISS

Tabular labradorite crystal, with numerous rounded inclusions of quartz and with large area of quartz just below it. Just to right and below a grain of garnet (black). A few grains of biotite are visible. x-nicols. x35.

FIGURE 2.—GALENA (WHITE) AND SPHALERITE (GRAY)

Note the distribution of fine specks of galena through the gangue (black) and inclusions of sphalerite in the galena. Reflected light. x120.

FIGURE 3.—PYRITIC ORE

Pyrite (white, high relief) corroded and rounded by galena (white to gray) and sphalerite (dark gray). Reflected light. x53.

FIGURE 4.—"PORPHYRITIC" ORE

Sphalerite and gangue black. Pyrrhotite shows initial stages of alteration to marcasite along grain boundaries. Pyrrhotite is slightly grayer than the galena which shows no alteration. Reflected light. x125.

FIGURE 5.—"EXSOLUTION" STRUCTURE

Grains of pyrrhotite (white) scattered through sphalerite (dark gray). Reflected light. x266.

FIGURE 6.—"EXSOLUTION" STRUCTURE

Veinlets and dots of pyrrhotite (white) in sphalerite (gray). Reflected light. x245.

PLATE 4

PARAGENESIS OF ORES

FIGURE 1.—CRYSTALLIZED MARCASITE

Crystallized marcasite (white) with pyrrhotite (gray) and an intermediate stage alteration preliminary to the crystallization of marcasite. Reflected light. x120.

FIGURE 2.—MARCASITE

Concentric structure of marcasite replacing pyrrhotite (dark gray) with the development of gangue-filled cavities. Reflected light. x120.

FIGURE 3.—MARCASITE

Well-crystallized marcasite, probably after pyrrhotite. Reflected light. x120.

FIGURE 4.—RELATIONS OF ORE AND GANGUE

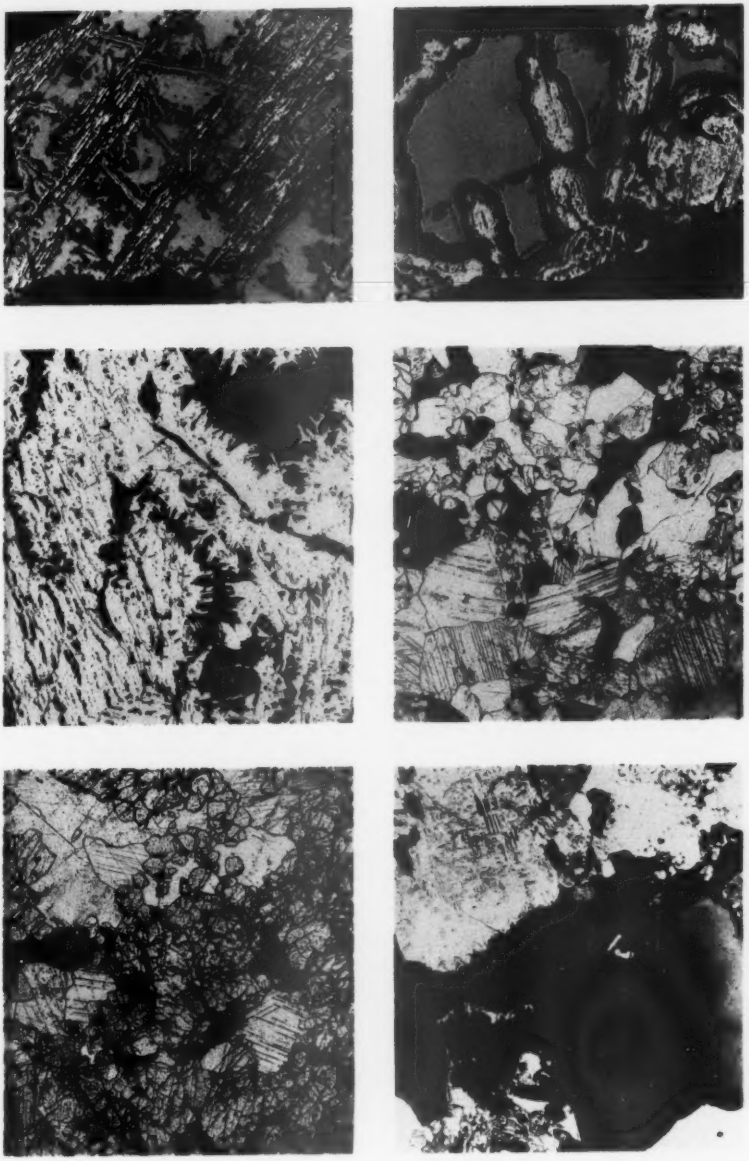
Sulphides (black) associated with silicates (fine-grained material) and relatively absent from the main carbonate areas (coarse-grained with strongly marked cleavage and twinning). Plain light. x21.

FIGURE 5.—RELATIONS OF ORE AND GANGUE

Galena, associated more closely with silicates (diopside) and spinel. Note partial rims of spinel (small octahedral grains with high relief) around carbonate grains (light-gray, low relief) and fine grains of galena scattered through the diopside. Plain light. x21.

FIGURE 6.—RELATIONS OF SPHALERITE TO PEGMATITE

Sphalerite veinlet in pegmatite, comprising quartz, myrmekite, and microcline-perthite. Note albitization of microcline along contact with sulphides. x-nicols. x35.



PARAGENESIS OF ORES

FIGURE 1. EXTERIOR OF TRUCK CARRYING AMPLIFYING EQUIPMENT
Seismo-pickups on ground to left of truck.

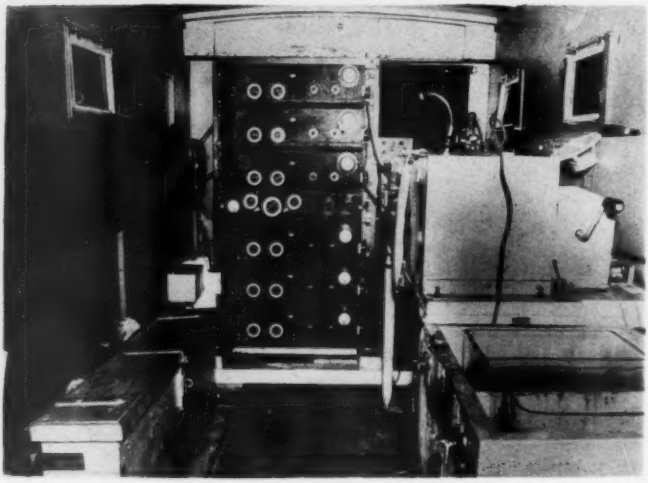


FIGURE 2. INTERIOR OF TRUCK WITH SEISMIC EQUIPMENT
Panel with six amplifiers in center; recording camera and telephone equipment to right center; developing and fixing tanks at lower right.

VIEWS OF TRUCK MOUNTED WITH SEISMIC EQUIPMENT

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 633-648, 2 PLS., 4 FIGS. MAY 1, 1941

SEISMIC STUDIES OF FLOORED INTRUSIVES IN WESTERN NEW HAMPSHIRE

BY F. C. KRUGER AND D. LINEHAN, S. J.

CONTENTS

Page

Abstract 633
Problem 635
Acknowledgments 637
Seismic surveying 637
  Principles of method 637
  Method adopted 639
  Application of method 639
    Equipment 639
    Limitations 639
Mascoma area 641
Bellows Falls area 642
  Geological description 642
  Velocity of direct waves 644
  Determinations of depth 644
    General statement 644
    Locality C 644
    Locality F 645
    Locality H 645
    Locality K 647
Summary and conclusions 647
Works to which reference is made 647

ILLUSTRATIONS

Figure Page

1. Generalized geological map of western New Hampshire 634
2. Velocity curves of seismic waves in Bethlehem gneiss at Bellows Falls localities. 636
3. Type of record of seismic reflections from lower contact of Bethlehem gneiss.. 638
4. Geological cross sections of Mascoma and Bellows Falls plutons 640

Plate Facing page

1. Views of truck mounted with seismic equipment 633
2. Geological map of Bellows Falls pluton 642

ABSTRACT

Geological mapping in western New Hampshire has delineated bodies of orthogneiss lying in structural basins of quartzite and schist. The object of the present investigation was to test the utility of seismic reflection methods for determining the depths of the lower contact of the gneiss at various points and to correlate these data with

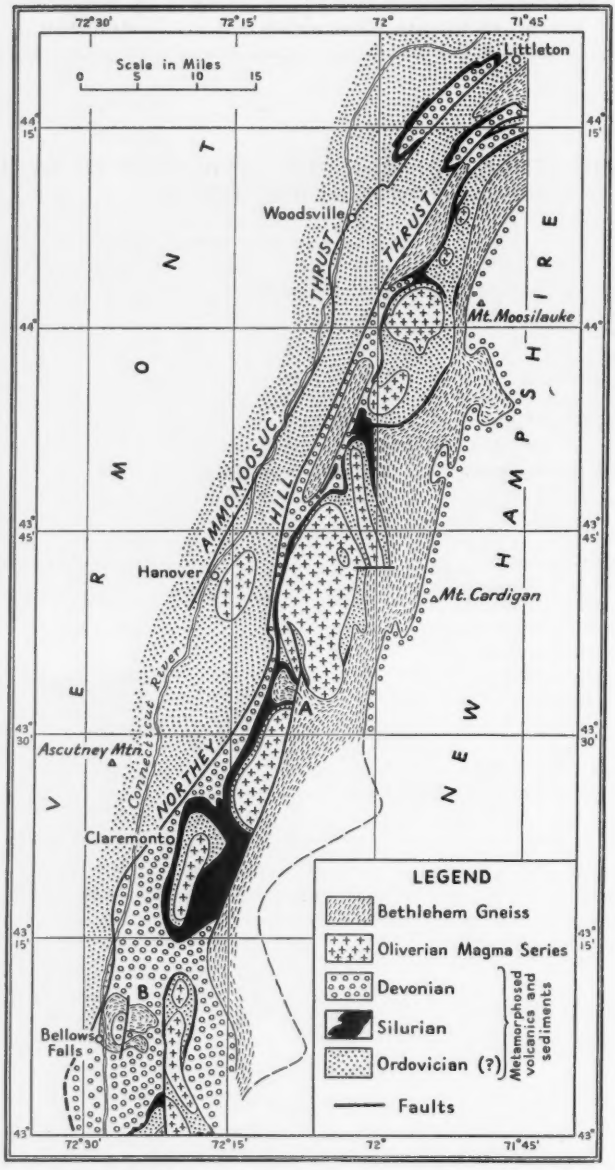


FIGURE 1.—*Generalized geological map of western New Hampshire*

A = Mascoma basin of Bethlehem gneiss; B = Bellows Falls basin of Bethlehem gneiss. Based upon published works of C. H. Hitchcock (1874, 1877, 1878), M. P. Billings (1937), K. Fowler-Lunn and L. Kingsley (1937), E. P. Kaiser (1938), C. A. Chapman (1939), J. B. Hadley and C. A. Chapman (1940), L. R. Page (1940), and upon unpublished works of M. P. Billings, C. A. Chapman, R. W. Chapman, and F. C. Kruger.

deductions made from geological observations at the surface. In one of the basins the seismic investigations were confined to one locality. In a second basin four localities were studied.

The surface velocity of the seismic waves in the orthogneiss varied from 11,600 feet per second to 25,250 feet per second, and at one locality it was possible to calculate from the reflected waves an effective average velocity of 17,000 feet per second. Reflections were obtained at all localities, except for one place where the records were illegible. In the basin where the studies were confined to one locality the discrepancy between the geological and seismic data was 20 per cent. In the second basin the average discrepancy was approximately 15 per cent.

The complete mapping of such a basin by seismic means would be difficult because of the nature of the terrain, which impedes the transportation of the equipment. The variation in the velocity of transmission of seismic surface waves introduces a factor of uncertainty. Moreover, the thickness of the glacial drift in many places is a complicating factor.

PROBLEM

Geological mapping in western New Hampshire has revealed metamorphosed sediments and volcanics which range in age from probable Ordovician to lower Devonian. Billings (1937) has assigned the intrusions to four magma series. The late Ordovician (?) Highlandcroft magma series and the Mississippian (?) White Mountain magma series are not involved in the present study. The Oliverian magma series, probably middle or late Devonian, occupies a series of intrusive domes, most of which are in the center of an anticline that can be followed for over 100 miles along the strike (Fig. 1). The Bethlehem gneiss, which belongs to the late Devonian (?) New Hampshire magma series, was injected for the most part concordantly with the bedding and schistosity. One pluton of the Bethlehem gneiss is a prominent structural unit east of the domes and is elongated parallel to the regional trend.

Masses of Bethlehem gneiss occupying structural basins are isolated west of the domes. Two of these outlying basins, at point A (Fig. 1) in the Mascoma quadrangle and at point B (Fig. 1) in the Bellows Falls quadrangle, have been studied by seismic methods in order to see how successfully geological and seismological data might be correlated.

The use of seismic methods for determining subsurface structure in the search for petroleum is well known. The results have more than justified the use of the method despite the large expense necessary to maintain a crew in the field. Practically all such surveys, however, are conducted in sedimentary rocks where the structure is relatively simple; moreover, the velocities are comparatively slow, and frequent calibration at wells is possible. The difficulties of studying steeply dipping beds (Heiland, 1936), are well known to the field observer and the computer. When employing seismic methods in areas where the structure is not only complicated but where rocks with high velocities are present, it is expected that the percentage of error will increase.

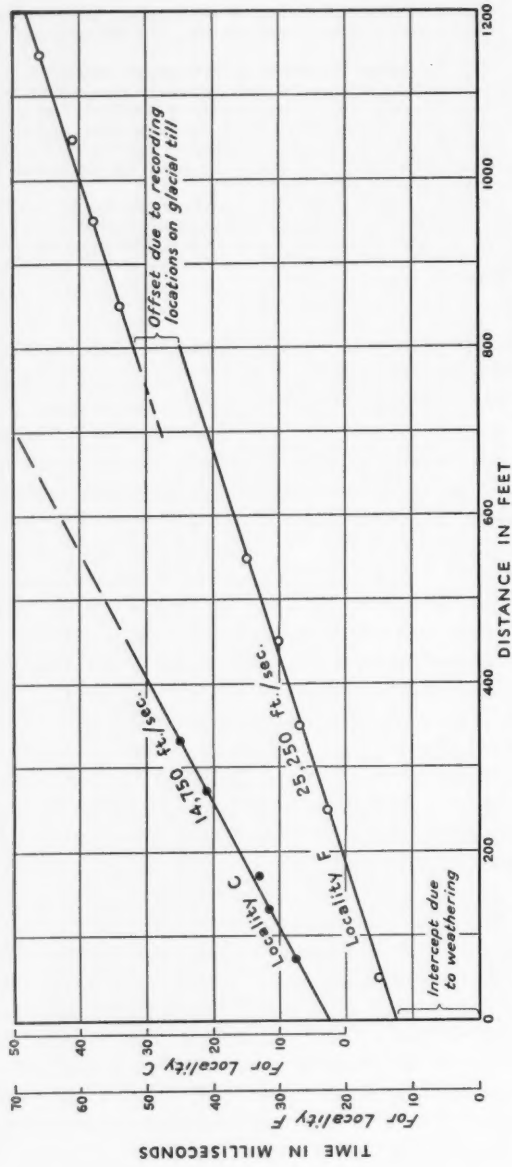


FIGURE 2.—*Velocity curves of seismic waves in Bethlehem gneiss at Bellows Falls localities*

Displacement of curve F at 800 feet is due to the depth of cover. Note that the curves for the two localities are plotted against separate vertical scales.

The geological field work on the Mascoma quadrangle by C. A. Chapman (1939) required the entire summers of 1935 and 1936, whereas that on the Bellows Falls quadrangle was done by F. C. Kruger during the summers of 1937 through 1940. The field observations were plotted on photostatic copies of the regular topographic sheets enlarged three times to the scale of 3 inches to a mile. In places of more complex geology or places where determination of location was difficult, pace and compass maps on scales of 200 to 880 feet to the inch were used. The seismological work was done by D. Linehan, S. J., in September 1938 and in August 1939.

ACKNOWLEDGMENTS

Sincere gratitude is due Professor Marland P. Billings and Professor L. Don Leet of Harvard University for their supervision of the field work and their guidance in the preparation of this paper. Both the seismological and geological studies were made possible by the Associates in Science of Harvard University. A loan of seismic equipment by the Humble Oil and Refining Company, Houston, Texas, and the advice and many courtesies of Mr. D. P. Carlton and Mr. D. Gardner of that company are gratefully acknowledged. Appreciation is due Mr. Trow and Mr. Newell of Alstead, New Hampshire, and Mr. Triboski of Charleston, New Hampshire, for consent to blast on their properties. The New England Power and Light Company furnished facilities for the necessary drilling in the Mascoma area.

SEISMIC SURVEYING

PRINCIPLES OF METHOD

The seismic reflection method is simple in principle. A dynamite blast at the shot point sets up vibrations that are received by instruments known as seismo-pickups which are placed at measured intervals away from the shot point. The vibrations received are amplified and recorded by the seismograph. The instant of the shot is recorded by an electrical timing device. The direct surface waves upon arrival at each of the seismo-pickups set up a vibration that continues until broken by the reception of seismic waves reflected from some horizon at depth. The reflected waves interfere and set up a new phase of vibration, thus causing a noticeable change in the character of the record. The elapsed time between the instant of shot and the time of arrival of the direct wave at each seismo-pickup gives the velocity of seismic waves for that rock. Figure 2 shows velocity curves of seismic waves. One half of the elapsed time between the instant of shot and the arrival time of the reflected wave at each seismo-pickup, multiplied by the velocity of the direct wave, makes a calculation of the depth to a reflecting horizon possible with

sufficient accuracy for these tests. Figure 3 shows the type of record of seismic reflections obtained. As successive reflected waves from deeper horizons are received, the vibrations recorded are changed by interference to show the effects of the newer reflections.

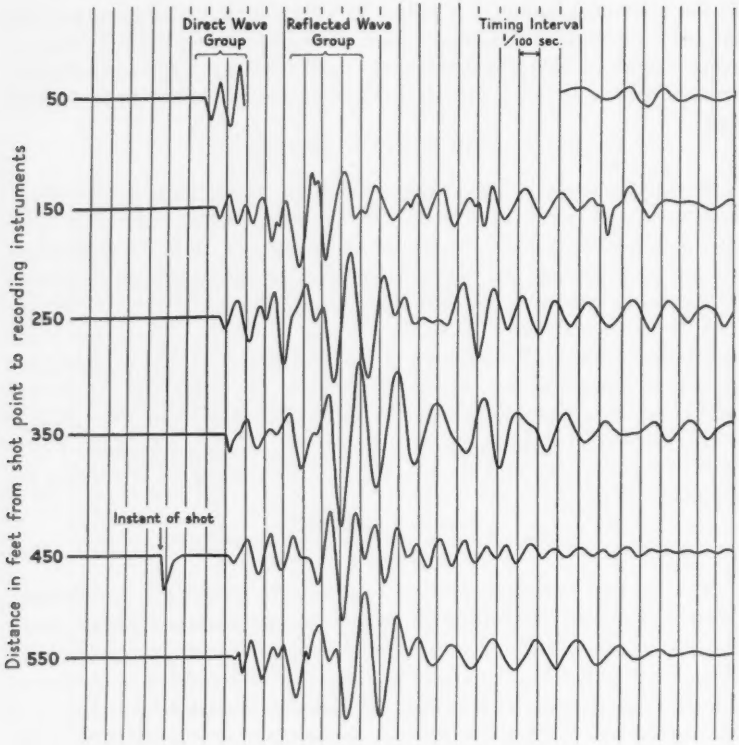


FIGURE 3.—*Type of record of seismic reflections from lower contact of Bethlehem gneiss*

Reflection is lost of the 50-foot trace due to proximity of explosion.

The velocity may also be determined in another way. What is known as the effective average velocity is computed by timing the reflections and plotting the square of the time against the square of the distance to the seismo-pickup.

The seismic refraction method requires a horizontal spread of the shot point and the recording instruments that is directly proportional to

the depth of the refracting horizon and inversely proportional to the velocities of seismic waves in the two rocks forming the refracting horizon. In this study it was not certain that the refracted waves from the lower layer of rock at any point were recorded. The profiles could not be extended to greater lengths with practicability; for this reason interpretations are based upon reflections only.

METHOD ADOPTED

To study these basins of Bethlehem gneiss the reflection method would be ideal if continuous profiles of the bottom contact could be maintained the entire distance across the bodies. The data thus obtained would be conclusive. The method actually adopted because of practical limitations made use of reflections from the bottom contact at selected convenient localities. This method, though not entirely satisfactory, presented data that permit some check on the geological structure sections.

APPLICATION OF METHOD

Equipment.—The bulky equipment of such a survey necessitates a truck (Pl. 1, fig. 1) for transportation of the seismo-pickups, connecting cables, timing and blasting wires, blasting box, telephone equipment, and batteries. Dynamite and digging tools are also carried. The truck is supplied with amplifiers for each of the seismo-pickups and a recording camera. The interior is designed as a darkroom including photographic tanks (Pl. 1, fig. 2).

Limitations.—The application of a seismic survey is limited by several factors, only some of which are geological. Transportation is perhaps the chief limitation of such a study. If, as in most cases, the amplifying and recording equipment is mounted in a truck, the working distance from the road is limited, of course, by the length of the channel wires from the seismo-pickups to the truck. If the equipment is portable, the working locations depend upon the physical capabilities of the field party. For the Mascoma locality portable equipment was used, but in the Bellows Falls locality the equipment used was mounted in a truck.

If the shot holes are to be in bedrock the seismic truck must be preceded by drilling equipment. Usually this limits the locations of shot points to short distances from the road because of the length of the air hose from the compressor. If the shot hole is to be in unconsolidated deposits, the location is determined by the ease of hand digging in the particular material.

Technically the individual outcrop should be massive and unjointed. It is desirable that the hole for blasting be 6 feet or more deep. The charge is tamped with water for each blast so that sufficient power is

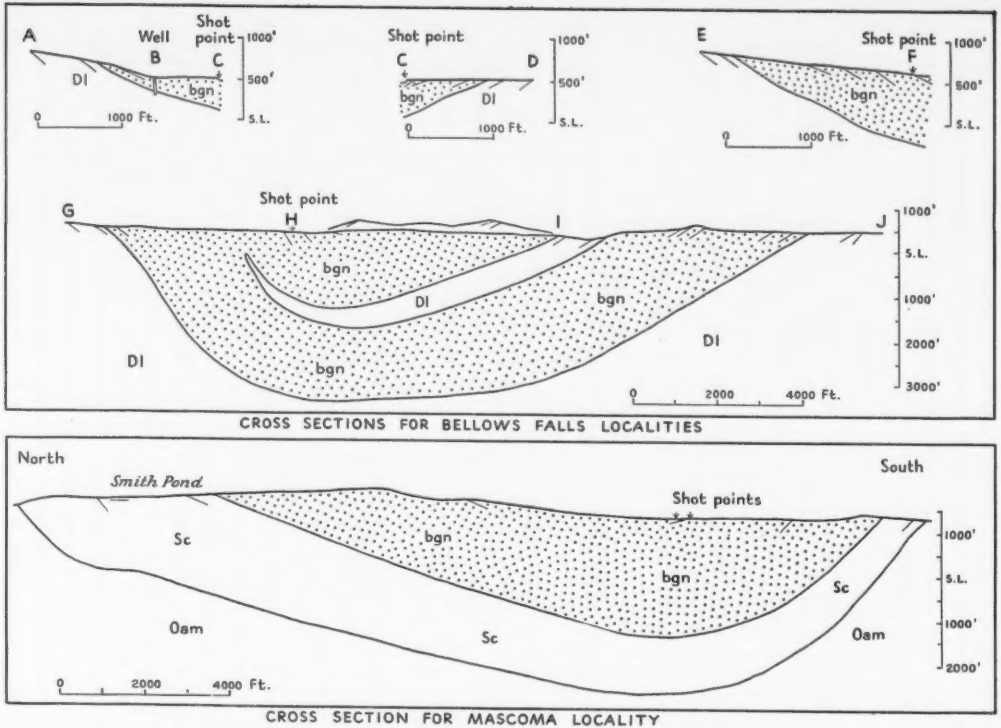


FIGURE 4.—*Geological cross sections of Mascoma and Bellows Falls plutons*

Bgn = Bethlehem gneiss, Dl = Devonian Littleton formation, Sc = Silurian Clough formation, Oam = Ordovician Ammonoosuc volcanics.

Representative geological structural data along the line of section are shown by short lines. *See* Plate 2 for location of sections in Bellows Falls area.

directed downward to set up recording waves, but at the same time expulsion of the water saves the hole for additional shots. Since many shots may be made at the same locality, a source of water for tamping should be near-by.

The seismo-pickups are connected along the cable at regular intervals which do not always permit the seismo-pickups to be placed on rock outcrops. Moreover, the outcrop must have a level surface in order that the seismo-pickups can stand nearly vertical. Trials were made in which the shot holes were in glacial drift and with the seismo-pickups placed on mineral soil; comparable results were obtained from these variations of the procedure. Induction from electrical transmission lines or electrified stock fences caused interference in the recording of the seismograph. Of course the deciding factor in the selection of any site for study was permission for blasting; obtaining this permission was troublesome only once.

MASCOMA AREA

Some of the Bethlehem gneiss in the Mascoma quadrangle which has been mapped by C. A. Chapman (1939) occupies a structural basin in the Silurian Clough formation. The body of gneiss is 2½ miles in diameter. The foliation of the gneiss and the bedding and schistosity of the metamorphosed sediments dip uniformly toward the center of the basin at angles from 35 to 55 degrees.

The Bethlehem gneiss is a medium-grained, biotite-rich rock ranging from granodiorite to quartz monzonite. The Clough formation consists of from 200 feet to 1200 feet of conglomerate, quartzite, and mica schist. Locally it is cut out by igneous intrusions. Though it is intimately contorted, the quartzitic nature of the Clough formation as contrasted with the Bethlehem gneiss provides an ideal discontinuity for seismic reflection methods.

The seismic surveys were first carried out on the Mascoma basin in September, 1938. Here two suitable areas were located near the center of the basin, but permission for blasting at one was refused. As a result of this, only two shot points about 300 feet apart were used for this preliminary work. At each point several holes for blasting were drilled and each hole could be used from five to seven times before becoming useless either by collapse or by the opening of fractures which allowed the tamping water to escape.

At this location the strike of the foliation of the gneiss is N. 10° E. dipping nearly vertically. The seismo-pickups were laid out in a traverse 450 feet long parallel to the strike of the gneiss. The geological cross sections for the Mascoma locality (Fig. 4) show the contact between the

Bethlehem gneiss and the top of the Clough formation to be 2650 feet below the surface or 1350 feet below sea level. The geological section also shows the contact between the bottom of the Clough formation and the top of the Ammonoosuc volcanics to be 3950 feet below the surface or 2650 feet below sea level.

The velocity of the direct wave in the gneiss obtained here was 11,600 feet per second. The contact between the Bethlehem gneiss and the Clough formation as determined by seismological methods is 2120 feet below the surface or 820 feet below sea level. The bottom of the Clough formation as determined seismically is 4150 feet below the surface or 2770 feet below sea level.

BELLOWS FALLS AREA

GEOLOGICAL DESCRIPTION

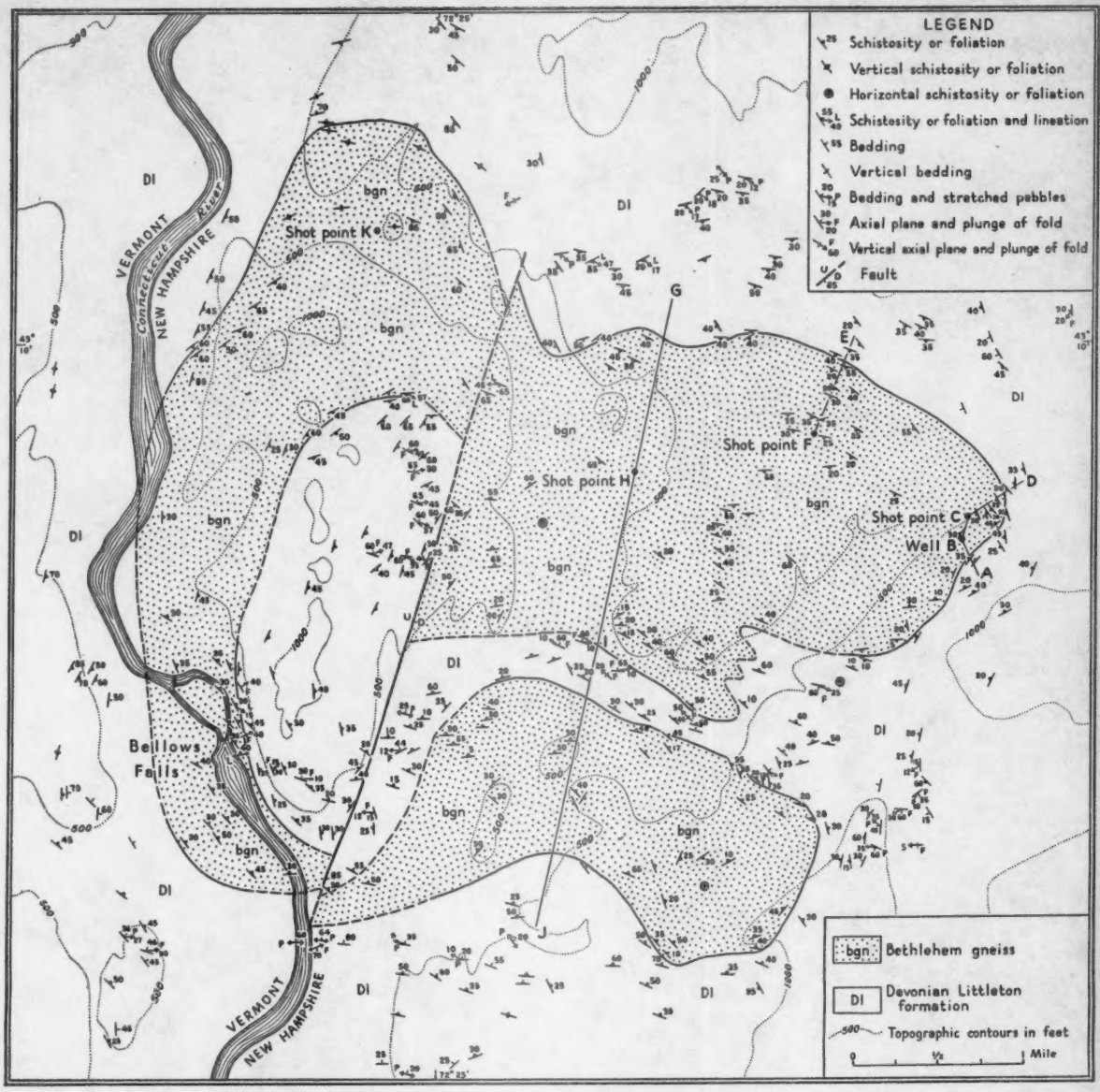
The structural basin in the Bellows Falls quadrangle (Pl. 2) is occupied by a thick sheet of Bethlehem gneiss which has been injected into the Devonian Littleton formation at a horizon 2500± feet above the top of the Clough formation. The intrusion of the gneiss took place during or after the last major folding of the region. Subsequent erosion has unroofed the gneiss. The enclosing metamorphosed sediments of the basin wrap conformably around the gneiss. The bedding, wherever observed, is parallel to the schistosity of the metamorphosed sediments, and both bedding and schistosity are parallel to the foliation of the gneiss.

Though the basal contact of the gneiss and the schist was not observed at any place, it is inferred that the contact is parallel to the concordant foliation and schistosity.

At the eastern edge of the gneiss (Pl. 2, point B) the log of an artesian well confirms the assumption that the contact is parallel to foliation and schistosity. The drill hole passed through the Bethlehem gneiss and obtained water from the underlying Littleton formation. The depth of the contact in the drill hole is consistent with the interpretation of the geology. The Bay State Artesian Well Company of Concord, New Hampshire, reports that the change of lithology from gneiss to schist was noticed in drilling operations 162 feet below the surface. The well is 200 feet deep and pumps 6 gallons per minute below 100 feet.

The inference that schistosity and foliation are in general parallel to the contacts of the Bethlehem gneiss was also found to be valid in the Moosilauke quadrangle by Billings (1937).

The foliation of the gneiss and the schistosity of the enclosing formations dip uniformly toward the center of the basin. In the southern part of the basin, the contact between the overlying Bethlehem gneiss and the Littleton formation dips 30 degrees to the north. Along the western



GEOLOGICAL MAP OF BELLOWS FALLS PLUTON

Based on representative geological data.

boundary the contact dips from 30 to 50 degrees toward the east. The dips at the northwest portion are vertical but decrease on the north edge of the basin to dips of 45 degrees to the south. At the eastern extremity of the basin the contact between the gneiss and the schist dips 30 degrees to the west.

The map (Pl. 2) shows a large mass of schist extending into the gneiss from the southeast. This body of schist forms a large septum or screen which intervenes between an upper portion of the gneiss to the northeast and a lower portion to the south and west. The attitude of the schist within the septum is entirely conformable with the schist that underlies the gneiss; both dip toward the center of the basin. The pattern of the minor folds, the attitude of the axial planes, and the plunge of fold axes all show that the septum of schist is a limb of a major fold (Fig. 4, sec. G-J). The persistence of the screen beneath the gneiss is impossible to deduce from surface structural data, but seismic evidence suggests extension as shown in the cross section. This interpretation shows the septum conformable with the underlying schist which forms the floor of the body of gneiss.

The Bethlehem gneiss has the average mineral composition of a quartz monzonite, and the grain size averages less than one millimeter in diameter. Though the rock may appear to be porphyritic, its typically granoblastic texture makes it wholly equigranular. The gneiss is unusually rich in biotite which gives it a noticeable foliation. In very favorable exposures a linear streaked arrangement of patches of biotite may be observed. Lineation may also be measured from the orientation of elongated inclusions.

The surrounding Littleton formation is for the most part composed of schist which is highly micaceous, containing varying amounts of quartzitic material. Discontinuous pure quartzite horizons are present but are quantitatively an unimportant part of the Littleton formation. A few pre-metamorphism amphibolitic volcanics are also interbedded with the schist. The rocks forming the floor of the basin are everywhere middle-grade types of the Littleton formation, containing such minerals as biotite, garnet, and staurolite. In general the septum of the Littleton formation that juts into the gneiss is composed of coarse-grained high-grade schist containing sillimanite and garnet.

The eastern half of the basin is the downthrow side of a normal fault with a displacement of approximately 1500± feet. No horizontal movement was noted. The fault plane where observed at one locality strikes N. 25° E. and dips approximately vertically. The body of gneiss in the Bellows Falls basin is from 3 to 5 miles in diameter and is accessible for seismological study in critical portions by roads.

VELOCITY OF DIRECT WAVES

More extensive studies were made of the structural basin in the Bellows Falls region in August 1939. The investigation was confined to four localities—points C, F, H, and K (Pl. 2).

Surprising differences of the velocity of the direct longitudinal waves in the Bethlehem gneiss were discovered, although the rock appeared to be petrographically uniform. At locality C, where the seismic profile was at right angles to the strike of the foliation, which at this place dips 25 degrees, the velocity was computed to be 14,750 feet per second. At locality F the profile was nearly parallel to the strike of the foliation of the gneiss which dips 25 degrees. The velocity here was 25,250 feet per second. At locality H the relationship of the seismic profile to the foliation of the gneiss was indeterminable because of the glacial drift. The record obtained at locality H, however, gave sufficient reflections to determine the effective average velocity as 17,000 feet per second. At locality K, the profile was at right angles to the vertical foliation. The velocity was 15,000 feet per second. The velocities of longitudinal waves in massive gneiss as given by Heiland (1940, p. 472, Table 52) are 14,896 to 24,606 feet per second.

The depth to the reflecting horizon is calculated by multiplying the velocity by one half the travel time of the reflected wave. It is apparent that some uncertainty is introduced into the calculations because of the range of the value of the velocity. (See Table 1.)

DETERMINATIONS OF DEPTH

General statement.—By means of the Busk (1929) method of the reconstruction of folds from surface geological data, the contact of the gneiss and the schist was projected in cross sections (Fig. 4) for the several shot points concerned. This method yields a somewhat larger figure for the depth of the gneiss than seismic sounding but checks remarkably with the actual well record. Considering the nature of the problem, the discrepancy between the results of the geological and seismic methods is reasonable. (See Table 1.)

Locality C.—At locality C the depth of the bottom of the gneiss can be calculated by geological methods in several ways. The section A-B (Fig. 4; Pl. 2) was constructed from the southern contact of the basin to the artesian well, at which point the depth of the contact was known to be 162 feet below the surface, that is, 338 feet above sea level. This actual depth was closely approximated by the figure of 170 feet below the surface or 330 feet above sea level, obtained by graphical construction. Continuing the section B-C (Fig. 4; Pl. 2) by using the surface data in

a projection from the bottom of the well, as the closest known datum, the depth of the contact at shot point C obtained by construction was 320 feet below the surface or 130 feet above sea level. Another section D-C (Fig. 4; Pl. 2) constructed from the eastern contact of the basin has good surface geological control and gives a depth figure of 400 feet below the surface or 50 feet above sea level at the same shot point C.

The strike of the seismic profile was east-west, approximately perpendicular to the foliation of the gneiss which strikes N. 20° W. and dips 25° SW. The profile was 350 feet long. For this profile the seismo-pickups were in each case placed on bedrock so that no corrections for cover were necessary. The shots were fired in a 6-foot drilled hole. The determination of the bottom of the gneiss by the seismic method at point C was 310 feet below the surface or 140 feet above sea level, using the observed velocity of 14,750 feet per second. The depth, obtained by using the effective average velocity for the Bethlehem gneiss was 357 feet below the surface or 93 feet above sea level.

Locality F.—Another traverse of shots was made on the northeastern flank of the body of gneiss at locality F where greater depths were expected and were encountered. The section E-F (Fig. 4; Pl. 2) based on geological observations shows the contact between the gneiss and the schist at shot point F to be 800 feet below the surface, that is, 200 feet below sea level.

The foliation of the gneiss at locality F strikes N. 70° W. and dips 25° SW. The seismic profile had a strike of west-northwest and was 1150 feet long. Here shots were fired both in bedrock at a depth of 6 feet and in glacial drift at depths of from 8 to 10 feet. The seismo-pickups were placed on bedrock wherever possible. Where the cover was shallow it was removed; otherwise the thickness of the cover was determined by seismic refraction or by measuring the displacement of the velocity curve due to this cover (Fig. 3). The determinations of the depth of the contact at locality F by seismic methods gave 780 feet below the surface or 180 feet below sea level using the observed velocity of 25,250 feet per second. If the calculation is made using the effective average velocity obtained for the Bethlehem gneiss, the depth is 525 feet below the surface or 75 feet above sea level.

Locality H.—An opportunity for a continuous seismic profile was afforded along a north-south road that traverses the gneiss. For a determination of much greater depth at point H (Fig 4; Pl. 2) near the center of the basin, larger amounts of charge were used, and the blasting was done in glacial drift. At this locality the widespread kame terraces prohibited good geological control. The greater thickness of cover made

TABLE 1.—*Results of geological and seismological studies*

	Mascoma locality	Bellows Falls locality			
		Point C	*Point F*	*Point H*	*Point K*
Depth of cover of glacial drift in feet	0	0	0–21	30–60	0
Length of profile in feet	450	350	1150	700	350
Strike of profile	N–S	E–W	WNW	N–S	N–S
Strike and dip of foliation	N. 10° E. Dip 90°	N. 20° W. Dip 25° SW.	N. 70° W. Dip 25° SW.	Unknown	N. 80° W. Dip 85° SW.
Surface velocity in feet per second	11,600	14,750	25,250	11,950	15,000
Depth in feet of gneiss using geological surface data	2650 (−1350)[a]	320 to 400[b] (+130)[a] (+50)[a]	800 (−200)[a]	1480[c] (−1050)[a]	
Depth in feet of a deeper geological horizon	3950[d] (−2650)[a]			3650[e] (−3220)[a]	
Depth in feet of gneiss using effective average velocity of 17,000 feet per second, computed from Δ^2 and t^2 of reflections at point H		357 (+93)[a]	525 (+75)[a]	1496 (−1066)[a]	
Depth in feet of gneiss using observed velocity	2120 (−820)[a]	310 (+140)[a]	780 (−180)[a]	1360 (−930)[a]	
Depth in feet of deeper reflections using observed velocity	4150[d] (−2770)[a]			30,000[e] (−29,570)[a]	

[a] Figure in parenthesis refers to sea level.

[b] Two sections based on different geological data give two different values of depth at Bellows Falls, locality C.

[c] Contact between the upper part of the gneiss and the septum of schist at Bellows Falls, locality H.

[d] Contact between the Clough formation and the Ammonoosuc volcanics, Mascoma locality.

[e] Contact between the lower part of the gneiss and the schist underlying the basin at Bellows Falls, locality H.

it necessary that the profile be run either side of the shot point H so that a reversal of profile could be obtained. The seismic profile was extended 350 feet north and south of the shot point H.

On the geological section G-J (Fig. 4; Pl. 2) the contact of the upper gneiss and the septum of schist is 1480 feet below the surface or 1050 feet

below sea level at point H. At the same locality the bottom contact of the lower gneiss and the schist is 3650 feet below the surface or 3220 feet below sea level.

Both of these discontinuities were registered by the seismograph at locality H. The contact of the septum is 1360 feet below the surface or 930 feet below sea level using the observed velocity of 11,950 feet per second for the seismic calculations. If the effective average velocity is used the depth obtained is 1496 feet below the surface or 1066 feet below sea level. A deeper seismic reflection at locality H, which may represent the bottom contact of the lower gneiss, was recorded at 30,000 feet below the surface or 29,570 feet below sea level using the observed velocity for calculation.

Locality K.—A measurement of the depth of the basin of gneiss was attempted in the northwestern part at locality K (Pl. 2). The foliation of the gneiss has an east-west strike and dips nearly vertically. The strike of the seismic profile was north-south. At this locality drilled shot holes did not permit the use of sufficient explosives so that no conclusive depth data were obtained. The velocity was measured as 15,000 feet per second. Moreover, point K was undoubtedly unfavorably located. It was equidistant from the steeply dipping walls of the basin of gneiss which might have been closer to the shot point than the inferred floor.

SUMMARY AND CONCLUSIONS

The data are summarized in Table 1. The seismic reflection method and the method of geological projection gave corroborative evidence of the depth of the structural basins of gneiss underlain by quartzite and by schist. The checks of the seismic method against the geological method are encouragingly within the expected limits of error of the two methods. At the Mascoma locality, and point C and point H of the Bellows Falls area, reasonably good checks were obtained; at locality F the check was not so good. The marked variations in velocity of the seismic waves in the gneiss were an unexpected complicating factor. When the problem was undertaken it was hoped that it would be feasible to chart both basins completely, but the limitations of the seismic apparatus in difficult terrain as well as the time and expense involved did not seem to justify expansion of the studies.

WORKS TO WHICH REFERENCE IS MADE

Billings, M. P. (1937) *Regional metamorphism of the Littleton-Moosilauke area, New Hampshire,* Geol. Soc. Am., Bull., vol. 48, p. 483-566.

Busk, H. G. (1929) *Earth flexures, their geometry,* Cambridge University Press.

Chapman, C. A. (1939) *Geology of the Mascoma quadrangle, New Hampshire,* Geol. Soc. Am., Bull., vol. 50, p. 127-180.

Fowler-Lunn, K., and Kingsley, L. (1937) *Geology of the Cardigan quadrangle, New Hampshire,* Geol. Soc. Am., Bull., vol. 48, p. 1363-1386.

Hadley, J. B., and Chapman, C. A. (1939) *The geology of Mt. Cube and Mascoma quadrangles, N. H.,* New Hampshire Planning and Development Commission.

Heiland, C. A. (1936) *Note on reflections from steeply dipping beds,* Geophysics, vol. 1, p. 257-270.

——— (1940) *Geophysical prospecting,* Prentice-Hall.

Hitchcock, C. H. (1874, 1877, 1878) *Geology of New Hampshire,* 3 volumes and atlas, Concord, N. H.

Kaiser, E. P. (1938) *Geology of the Lebanon granite, Hanover, N. H.,* Am. Jour. Sci., 5th ser., vol. 36, p. 107-136.

Page, L. R. (1940) *Geological map and structure sections of the Rumney quadrangle, New Hampshire,* State of New Hampshire, Highway Department.

Harvard University, Cambridge, Mass.; Weston College, Weston, Mass.
Manuscript received by the Secretary of the Society, January 14, 1941.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 649-666, 2 FIGS. MAY 1, 1941

ROOTS OF THE TACONIC FAULT IN WEST-CENTRAL VERMONT

BY H. E. HAWKES, JR.

CONTENTS

	Page
Abstract	650
Introduction	650
Taconic thrust fault	650
Problem	652
Acknowledgments	652
Stratigraphy	652
General considerations	652
Quartz schists	654
Pico Peak series	654
Plymouth Union series	654
Sericite schists	656
Pinney Hollow schist	656
Ottauquechee group	656
Thickness	656
Age	657
Minor structures	657
Foliation and lineation	657
Slip cleavage	658
Late strain systems	658
Major structure	658
Folding	658
Taconic thrust fault	660
General statement	660
Similarity of Taconic sequence and sericite schists	660
Sedimentary contact at base of sericite schists	661
Truncation of structures	662
Consistent structural pattern	663
Absence of Cheshire quartzite on east side of range	663
Homoclinal structure of Green Mountains	664
Contrast of Green Mountain and Valley structures	664
Review of critical areas	664
Summary and conclusions	665
Works to which reference is made	666

ILLUSTRATIONS

Figure	Page
1. Geologic structure of western New England	651
2. Plan and generalized section showing relations of Green Mountain, Taconic, and Valley sequences	655

ABSTRACT

A detailed field examination of an area between Plymouth and Rochester, Vermont, on the east side of the Green Mountain range was made in an attempt to locate the root zone of the Taconic thrust fault. Evidence was found that the roots of the thrust do not outcrop in the area studied. The discordant relationships of the Green Mountain sequence to the marbles of the Rutland valley suggest that the fault lies a short distance east of the marble belt. The Green Mountain sequence in this latitude appears to form an eastward-dipping homocline rather than an anticline as previously thought. A minimum figure of 14,000 feet is estimated for the stratigraphic thickness of the Green Mountain sequence.

INTRODUCTION

TACONIC THRUST FAULT

Investigators of the structure of the Taconic Mountains of western Vermont and Massachusetts have found evidence for a gigantic thrust fault, the Taconic thrust, along which the Cambro-Ordovician argillaceous sediments of the Taconic Mountains have been shoved many miles westward over the contemporaneous calcareous rocks of the Valley sequence. The most significant evidence in favor of the thrust hypothesis is based upon a consideration of depositional facies. Generally accepted concepts of sedimentary environments will not permit an abrupt lateral transition from an exclusively calcareous to an exclusively argillaceous type of deposition. It is necessary to postulate that these different types of sediments were laid down in separate basins or different parts of the same basin wherein for some reason the environments of deposition were different. The only explanation which adequately accounts for the present intimate association of contemporaneous clay and lime sediments is horizontal faulting on a grand scale.

The argillaceous rocks of the Taconic thrust sheet lie in a huge thrust outlier, or klippe, which extends from west-central Vermont southward probably as far as the Catskills (Fig. 1). No physical connection exists today between this klippe and the roots of the thrust, as erosion has exposed the rocks beneath the thrust sheet on all sides. Since no structures which could possibly contain the roots of the thrust occur north, west, or south of the klippe, it has been concluded that the fault plane plunges back into the depths to the east of the Taconic Range, in the heart of the Green Mountains of western New England.

In the opinion of the present author, previous attempts to locate the eastern roots of the Taconic thrust fault have not yielded convincing results. Prindle and Knopf (1932) in the Taconic quadrangle of western Massachusetts suggest that it outcrops on the west slopes of Hoosac Mountain, but this area is so complicated that the evidence is not very clearly defined. Keith (1932), in Plymouth township of central Vermont, noted the striking lithologic similarity between the green schists east of

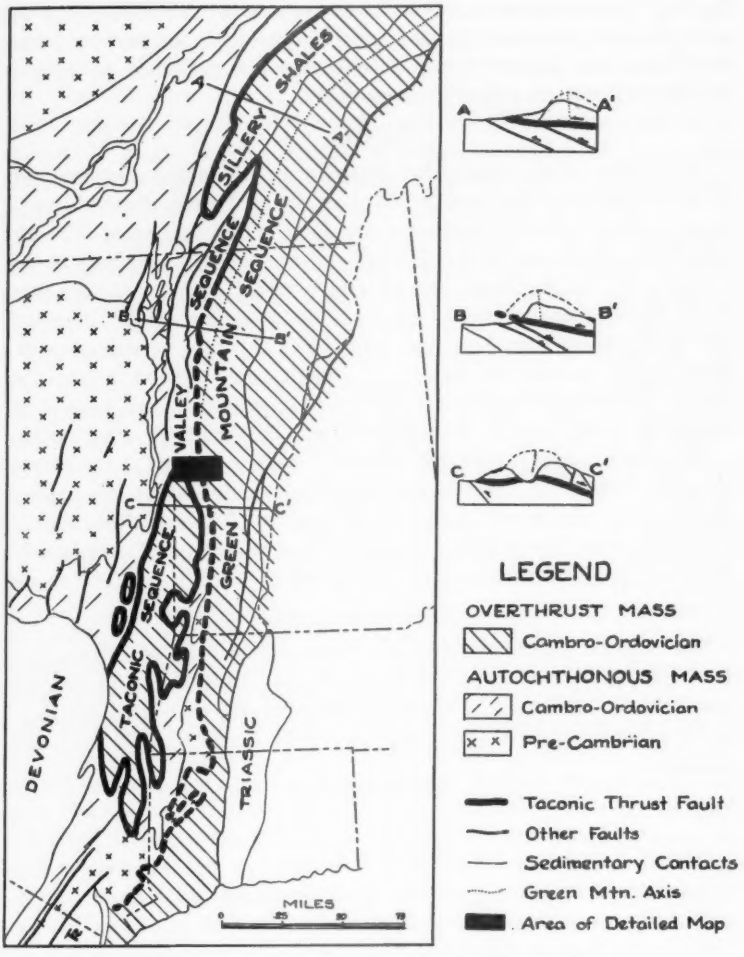


FIGURE 1.—*Geologic structure of western New England*
After Kay (1937).

the Green Mountain range and the slates of the Taconic sequence and assumed that they were taxonomic equivalents. He considered the quartzites and dolomitic marbles beneath the schists of Plymouth township to be the exposure of the autochthonous Rutland marbles on the eastern limb of the Green Mountain anticline. From this he concluded

that the Taconic thrust outcrops at the contact of the schists and the quartzites and marbles. He mentioned that he had also found evidence for a structural discordance at this contact but published no map or detailed field data to support this conclusion.

PROBLEM

The problem of the present investigation was the detailed examination of an area which might contain the roots of the Taconic thrust in order to prove the presence or absence of the fault within that area. The terrane between Plymouth and Rochester, Vermont, (Fig. 1) was chosen because: (1) Keith (1932) had already reported evidence for the roots of the thrust from that area; (2) Perry's (1928) earlier work in Plymouth and Bridgewater townships had established the broad outlines of the stratigraphy; and (3) the structures of the Taconic fault at the northern end of the klippe, 20 miles northwest of Plymouth, had been worked out in greater detail than in any other locality (Kaiser, 1938; Cady, 1938).

The present paper is a condensation of a thesis submitted to the Massachusetts Institute of Technology in partial fulfillment of the requirements for the degree of Doctor of Philosophy. The complete report is obtainable from the Institute library. Four and one half months were spent in the field during the summer and fall of 1939.

ACKNOWLEDGMENTS

The writer wishes to express his indebtedness to Prof. W. J. Mead and Prof. H. W. Fairbairn, of the Massachusetts Institute of Technology, and Prof. G. M. Kay, of Columbia University, for their kind assistance and criticism both in the field and during the preparation of the manuscript.

STRATIGRAPHY

GENERAL CONSIDERATIONS

Perry's mapping of Plymouth and Bridgewater townships east of the Green Mountain range revealed two major lithologic groups: (1) the underlying quartzites, quartz schists, dolomitic marbles, and gneisses of the Green Mountain axis; and (2) the overlying argillaceous schists and quartzites. He divided the lower group into three series—the Mt. Holly gneiss of "Archean" age, the Mendon series of "Algonkian" age, and a younger group of less metamorphosed quartzites and dolomitic marbles of "older Cambrian" age. The upper group was subdivided into four natural classifications, the lowest Pinney Hollow green sericite schist, the Ottauquechee group of black phyllites and quartzites, the Bethel green schist, and the Missisquoi group of green schists and argillaceous quart-

zites. He considered that the formations of this group were of younger Cambrian age, since they underlie with apparent unconformity the Ordovician Memphremagog group of central Vermont. No fossils have yet been identified from any of these formations, so that Perry's age classification is open to question. The rocks of both groups are commonly referred to as the Green Mountain sequence, as distinguished from the autochthonous Valley sequence of the Champlain and Rutland valleys, and the overthrust Taconic sequence of the Taconic Mountains.

On the average, the rocks of Plymouth and Bridgewater townships strike north-south and dip to the east. Perry, following Whittle (1894), was of the opinion that they lie on the eastern flank of a major anticline, the axis of which coincides with the crest of the Green Mountains.

In tracing the horizons of Plymouth township north to Rochester, the present author found reason to revise the stratigraphic nomenclature of the lower part of the Green Mountain sequence. The distinguishing features of the three lower groups appear to fade out toward the north. In the absence of any evidence for unconformities throughout the section, the writer is of the opinion that these groups all belong to the same continuous sedimentary series, with merely local variations in lithology. Perry's use of the term "Mendon series" was based on the assumption that the quartz schists of Plymouth are equivalent to the Mendon series from the type locality east of Rutland (Whittle, 1894) and that they lie on opposite flanks of the Green Mountain anticline. Since there is not only no lithologic correspondence between the two, but also definite evidence that the structure of the Green Mountains is not anticlinal in this latitude, the term Mendon series becomes confusing when applied to the rocks of Plymouth township. The Mt. Holly series was originally defined as those rocks which lie beneath the Mendon series. Since the latter has not been identified east of the type locality, it seems best also to discard the term "Mt. Holly" from discussions of the eastern flanks of the Green Mountains.

The present author proposes the general term "quartz schists" to include all the rocks of the Green Mountain core which lie beneath the Pinney Hollow schist on the east side of the range. He suggests that the lower part, which Perry designated "Mt. Holly," be called the "Pico Peak series," and the upper part, including Perry's Mendon series and Older Cambrian group, be called the "Plymouth Union series." It should be emphasized that these are intended as purely local names, which may prove to be not applicable outside the area herein described. The formations above the Plymouth Union series and beneath the Ordovician Memphremagog group are grouped together as "sericite schists."

The proposed nomenclature for the rocks beneath the Pinney Hollow schist is as follows:

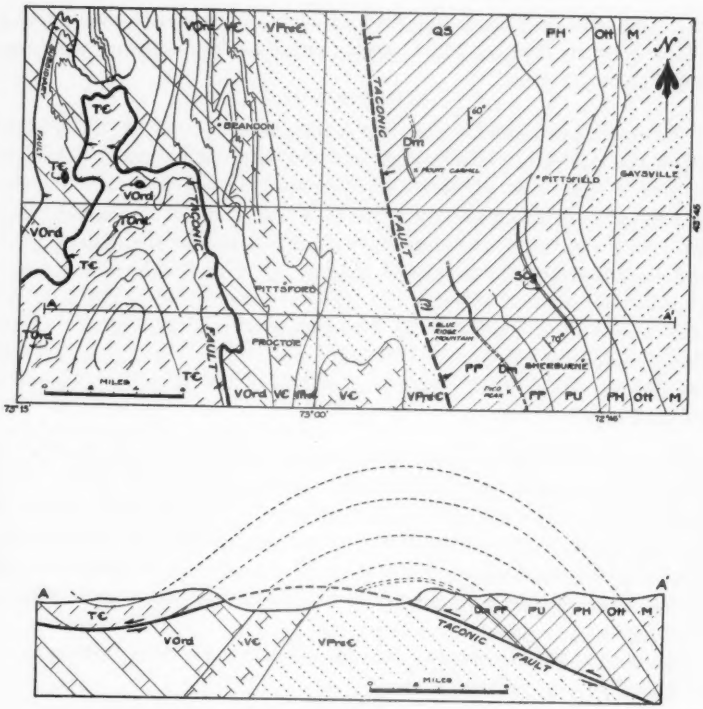
Proposed Nomenclature	Perry's Nomenclature	Lithology
Plymouth Union series	Older Cambrian group	Quartzites, dolomitic marbles, quartz schists
	Mendon series	
Pico Peak series	Mt. Holly series	Quartz schists, dolomitic marble, sedimentary gneisses

QUARTZ SCHISTS

Pico Peak series.—The Pico Peak series is undifferentiated quartz schists, quartzites, and quartzose gneisses. While these rock types make up 98 per cent of the stratigraphic column in this series, variations in grade and intensity of metamorphism over short distances render them useless as marker horizons. A distinctive bed of pure dolomitic marble, 100 feet or more in thickness, was the only traceable bed in the series. The most satisfactory exposures of this horizon are in the valley east of Blue Ridge Mountain, but an outcrop of similar dolomite occurs on the north slope of Pico Peak which is thought to be part of the same formation. The trend of this dolomite bed closely parallels the trend of the structure in the less obscure terrane to the east and is believed to be part of the same sedimentary sequence.

Plymouth Union series.—The Plymouth Union series is characterized by quartz schists, quartzites, quartz conglomerates, and dolomitic marbles. The basal contact with the Pico Peak series is marked by a clean massive quartzite. Near Plymouth, the upper part of the series contains several 50-foot beds of massive dolomite which finger out toward the north. Near the top of the Plymouth Union series north of Sherburne, a well-defined conglomerate, the Sherburne conglomerate originally described by Richardson and Maynard (1938), may be traced for 4 miles along the strike. A 600-foot transition zone between this series and the sericite schists apparently reflects a change of depositional environment from near-shore to off-shore conditions. It is composed of rocks intermediate in composition between the quartz schists and dolomites below and the sericite schists above.

North of North Sherburne, the distinctive horizon markers of the quartz schists lose their identity as a result of intense secondary deforma-



EXPLANATION

GREEN MOUNTAIN SEQUENCE

- M — Bethel Schist & Missisquoi Group
- Ott — Ottauquechee Group
- PH — Pinney Hollow Schist
- SCg — Sherburne Conglomerate
- Dm — Dolomitic marble
- Undifferentiated Quartz Schists (QS)
  Plymouth Union Series (PU)
  Pico Peak Series (PP)

TACONIC SEQUENCE

- TOrd — Ordovician slates
- TC — Cambrian slates

VALLEY SEQUENCE

- VOrd — Ordovician marbles
- VC — Cambrian marbles
- VPreC — Pre-Cambrian

FIGURE 2.—*Plan and generalized section showing relations of Green Mountain, Taconic, and Valley sequences*

Rutland Valley from Bain (Longwell, 1933); Taconic Sequence from Kaiser (personal communication); Brandon Valley from Cady (personal communication).

tion. The only horizon in this area which was mapped in detail was a 75-foot dolomite bed on Mt. Carmel, which could be traced with certainty for only a mile.

SERICITE SCHISTS

Pinney Hollow schist.—The predominant rock type of the Pinney Hollow is a pale-green sericite-quartz schist. Except for minor persistent horizons of graphitic phyllite up to a few tens of feet in thickness, the formation is exceptionally uniform in composition and appearance. North of Pittsfield two apparently concordant sheets of brilliant green rock composed of epidote, albite, chlorite or green biotite, calcite, and magnetite appear and widen to the north to a maximum of 600 feet at a point a few miles southwest of Rochester. This is interpreted as the alteration product of a lava flow or shallow intrusion. The calcite marbles described by Dale (1915) from Rochester and Hancock are a phase of this rock type.

Ottauquechee group.—The Ottauquechee group is characterized by its black carbonaceous content, which distinguishes it from the green schists of the Pinney Hollow. The dominant rock types are black, graphitic phyllite, massive-gray quartzite, and finely interlaminated pale-gray quartzite and schist. This group has been traced from Ludlow township (Perry, 1928) to Stowe (Jahns, personal communication), a total distance of 70 miles.

The present detailed survey was not extended above the Ottauquechee group.

THICKNESS

The rocks of the Green Mountain sequence in the area between Plymouth and Rochester strike in general north-south and appear to dip at an average angle of about 60° E. No evidence was found for either large-scale or small-scale repetition, though Jahns (personal communication) reports that he has found good evidence for repetition by small isoclinal folds in the Barre, Vermont, quadrangle sufficient to give an apparent thickness two and one half times the true thickness. The width of exposure of the Green Mountain sequence from the oldest known members to the base of the Memphremagog group varies from 50,000 to 85,000 feet. If we accept the thickening ratio of two and one half to one suggested by Jahns, the true stratigraphic thickness of the section varies between 14,000 and 24,000 feet. On the other hand, if the apparent dip represents the true dip, the total thickness is between 40,000 to 70,000 feet. Neither estimate includes the great thickness of the overlying Memphremagog group. While these exact figures may be questioned, some idea may be gained from this of the magnitude of the Green Moun-

tain sequence, which was apparently laid down in one of the major geosynclinical troughs of the North American continent.

AGE

Very little is known regarding the age of the Green Mountain sequence of Vermont, as no fossils have yet been reported from any of the formations herein described. The probable extension of the sericite schists into Canada appears to lie beneath Ordovician Sillery shales. The absence of evidence for an outstanding unconformity in the section reduces the possibility of a pre-Cambrian age, so that it seems reasonable that the sericite schists are Cambrian. The position of the sericite schists beneath the Ordovician Memphremagog group is also consistent with a Cambrian age. Beyond this, it is impossible to make any definite statements regarding the age of the schists. It is interesting that the presumed equivalence of the Taconic slates and the sericite schists also requires a Cambrian age for the latter.

A pre-Cambrian age has been generally held for the rocks of the core of the Green Mountains. The best supporting evidence for this is an unconformity between the Lower Cambrian Cheshire quartzite and the underlying formations at Sucker Brook near Lake Dunmore on the west side of the range. As discussed below, there is evidence that the quartz schists are not equivalent to the formations beneath the Cheshire quartzite and therefore may be of an entirely different age. On the east side of the range, Richardson and Maynard (1938) considered the Sherburne conglomerate to be a basal conglomerate of Cambrian schists, overlying pre-Cambrian gneisses. The present study has shown that the Sherburne conglomerate lies almost 1000 feet beneath the base of the sericite schists, that its flood-plain origin is questionable, and that the underlying granite gneisses are the result of granitization after the conglomerate was already in place. For these reasons, the Sherburne conglomerate is not held to be a good criterion for the pre-Cambrian age of the underlying rocks. All the evidence so far found indicates that the quartz schists belong to an earlier part of the same continuous depositional sequence as the sericite schists.

MINOR STRUCTURES

FOLIATION AND LINEATION

In almost all cases where both bedding and foliation were observed in the schists of the Green Mountain sequence they appeared to be exactly parallel. The strike is generally north-south with a steep easterly dip. The foliation is relatively coarse and irregular and is associated with a linear structure which plunges steeply north in the plane of the foliation. In all the slides studied there was a pronounced orientation of quartz

axes in a girdle about the lineation. The writer believes that the lineation is not related to the stresses which produced the major folding but is the result of later deformation in which horizontal shearing predominated. The present coarse foliation is thought to be the result of an intensification of early bedding-foliation by these shear stresses.

SLIP CLEAVAGE

Later slip cleavage or "strain-slip" cleavage cuts the foliation. It appears commonly in systems of closely spaced vertical shear fractures which strike north-south and is generally associated with similarly oriented zones of harmonic plications, where the east side has moved north and up relative to the west. Apparently no significant movement has taken place along the slip cleavage planes themselves. This phase of the deformation served to destroy the quartz fabric associated with the earlier lineation generally without developing a new orientation of quartz axes.

LATE STRAIN SYSTEMS

Deformational twinning on the $(02\bar{2}1)$ plane in dolomite crystals occurs in many of the dolomite-bearing rocks of the area. Unfortunately, the stress-strain relationships of dolomite twinning have not been investigated in the laboratory. If a relationship similar to that observed in calcite were assumed, however, dolomite glide lines conform statistically to a strain system oriented uniformly with respect to geographic co-ordinates rather than to the local attitude of the foliation or slip cleavage. This strongly suggests that the dolomite twinning reflects a stress late in the tectonic history, when the early foliation and slip cleavage structures had been annealed to such an extent that the rock failed as an isotropic mass. The characteristics of dolomite twinning is described in greater detail in a forthcoming paper by Fairbairn and Hawkes.

MAJOR STRUCTURE

FOLDING

Some of the earliest investigators in Vermont (Adams, 1845; Hitchcock, 1861; Whittle, 1894) considered the major structure of the Green Mountains to be anticlinal. This view was held largely because at that time the metamorphosed quartz schists were thought to be pre-Cambrian in age, flanked by less metamorphosed, and hence younger sediments. Since the pre-Cambrian age of the quartz schists had been seriously questioned, this argument loses some of its significance.

More reliable evidence for an anticlinal structure comes from a consideration of the Green Mountain structures on the extension of the range into Canada, where the metamorphism is apparently less intense. Fair-

bairn (1932, p. 71-75) and Clark (1934, p. 15) describe an anticlinal structure in the Sutton Mountains. Cooke (1937, p. 16) in the Thetford district of Quebec found a similar anticlinal pattern in the core of the range. It should be emphasized that the continuation of this anticlinal structure to central and southern Vermont in the opinion of the present author has not been adequately established by field work. The conclusions of Whittle (1894) and Foye (1918) appear to be established on the assumption that the structure is anticlinal without any clear-cut local evidence for such a structure.

Within the quartz schists of the area between Plymouth and Rochester, the apparent dip of the beds is consistently 30° to 80° E. No evidence was found for arching of the bedding across the range similar to that which occurs farther north. The minor structures, such as drag folds and foliation, do not appear to be associated with the major folding and cannot be used in working out the major fold pattern. Furthermore, no sedimentary criteria for tops and bottoms were found either in the present survey or in reports of earlier work in the core of the Green Mountains. Without more evidence, it is impossible to say whether the structure of the Green Mountain axis is an isoclinal anticline, an isoclinal syncline, or an eastward-dipping homocline. The only remaining approach to the problem is a consideration of the possible recurrence of distinctive horizons or stratigraphic sequences such as might be found on the opposite limbs of a major fold.

Several horizons in the latitude of Sherburne can be followed for considerable distances along the strike (Fig. 2). The contact of the quartz and the sericite schists, the Sherburne conglomerate, the basal quartzite of the Plymouth Union series, and the dolomite of the Pico Peak series are all relatively good marker horizons and are all closely parallel. As far west as the base of Blue Ridge Mountain no recurrence of any distinctive bed or of any part of the stratigraphic sequence was observed.

The dolomite of the Pico Peak series is a thick bed which retains its individuality through extreme variations in metamorphic intensity. If this dolomite were on the western limb of an anticline, its exposure on the eastern limb must lie somewhere to the east. No corresponding bed was found there, however. In view of the thickness and the persistence of this bed, it seems unlikely that it lenses out in an east-west direction to such an extent that it would be unrecognizable on the other limb of a fold structure. Therefore, we are led to the conclusion that this horizon lies on the eastern flank of the anticline, together with all the superjacent formations. If this is the case, the anticlinal axis lies still farther west.

Foyles' (1938) work on Blue Ridge Mountain revealed only one outcrop of dolomite between the dolomite described above and the Rutland valley. Because of poor exposures he found it difficult to estimate accurately the thickness or strike of this bed so that no definite correlation with the dolomite of the Pico Peak series seems justified.

This evidence suggests that the structure of the quartz schist immediately north of Pico Peak is homoclinal rather than anticlinal, at least as far west as Blue Ridge Mountain. This conclusion is of utmost importance in considering the structural relationships of the Green Mountain sequence to the marbles of the Rutland valley, as will be shown.

Unfortunately, a similar analysis of the dolomite of Mt. Carmel cannot be made at present. Although considerable reconnaissance work was done east of Mt. Carmel, it was not sufficiently detailed to rule out the possibility of the recurrence of a similar horizon to the east. Nevertheless, the fact that none was found suggests that this dolomite also lies on the eastern flank of the anticline and that a homoclinal structure prevails in that area as well.

TACONIC THRUST FAULT

General statement.—The main problem of the present investigation was the question of whether any structures in the Plymouth-Rochester area can be interpreted as the roots of the Taconic thrust fault.

At the northern tip of the thrust klippe in the vicinity of Brandon, the nature of the fault contact has been investigated in considerable detail by Cady and Kaiser. They have observed the following structural features associated with the sole of the thrust in that area: (1) abrupt change in lithology from an argillaceous to a calcareous facies; (2) truncation of contacts in both sequences; (3) a notable contrast in structural patterns in the two sequences; (4) the presence of "fensters," or thrust windows of Valley rocks showing through the main mass of the Taconic sequence; (5) general absence of minor structures associated with the thrusting; and (6) some associated subsidiary thrusting. Cady's and Kaiser's maps illustrating these structures have been reproduced in Figure 2. They found no contrast in metamorphic grade between the two sequences, which suggests that the thrusting occurred prior to metamorphism.

With the typical features of the Taconic fault near Brandon as a guide, the author carefully examined the structures of the Green Mountain sequence between Plymouth and Rochester. No field evidence was found for placing the roots of the Taconic thrust in the area studied.

Similarity of Taconic sequence and sericite schists.—Keith (1932) has pointed out the striking lithologic similarity between the green and black

argillaceous schists of the Pinney Hollow and Ottauquechee formations, and the green and black argillaceous slates of the Taconic range. Fossil data place the latter predominantly in the Cambrian, and the available evidence favors a similar age for the sericite schists. Until some evidence to the contrary is presented, it is reasonable to accept the taxonomic equivalence of these groups as a working hypothesis. From this, it follows that the sericite schists belong to the same overthrust sedimentary sequence as the Taconic rocks and that the roots of the Taconic fault therefore outcrop between the base of the sericite schists and the Valley marbles.

Sedimentary contact at base of sericite schists.—The feature of the geology which Keith (1932, p. 404) emphasized was the contrast in sedimentary facies between the sericite schists and the quartz schists, suggestive of the contrast between the Taconic and Valley sequences respectively. This was his main argument for placing the fault at the basal contact of the sericite schists. Where this contact is well exposed, as in the beds of tributary brooks north of Sherburne, it is seen to be gradational across a thickness of 600 feet. The present author found that this transition zone appears to have none of the features of the sharp contact of the Taconic slates near Brandon but appears to represent merely a normal change in depositional environment. It might be argued that this is not a true sedimentary transition but that it represents a zone of isoclinal infolding of the schists with the quartzites, and that the original contact before deformation was as sharply defined as the thrust contact near Brandon. Important features to bear in mind in this connection are: (1) that the transition zone is persistent, both in thickness and lithology, and (2) that it is not a composite of typical Pinney Hollow schist and Plymouth Union quartz schists but has characteristics of its own. The association of gray marbles with the black phyllites of the transition zone is unique. No pure calcareous sediments were found in the sericite schists, and where they occur in the quartz schists they are light-colored and are associated with quartzites rather than carbonaceous and argillaceous rocks. Furthermore, the sericite-quartz schists of the transition zone appear to be midway in mineral composition between the typical Pinney Hollow argillaceous schist and the more highly siliceous quartz schists of the underlying series. The evidence seems strongly to favor the thesis that this contact reflects merely a change in sedimentation from a sand-lime facies to a clay facies.

Another part of Keith's argument for placing the fault at the base of the Pinney Hollow is that the upper horizons of the Plymouth Union series appear to be truncated by the contact. Perry's map of Plymouth

township shows the structure following the valley north of Plymouth Union, so that the entire section of his "Older Cambrian" series is squeezed into the width of the Ottauquechee Valley north of West Bridgewater. As outcrops are poor in the valley bottom, Keith considered it just as likely that the formations, instead of thinning, disappear beneath the Pinney Hollow. The field evidence was re-examined by the writer, who found that the formations do not follow the valley but strike parallel to the base of the Pinney Hollow up the hillside west of the valley. While some of the conspicuous calcareous horizons pinch out, the apparent thickness of the section remains constant.

Truncation of structures.—Wherever satisfactory marker horizons were found in the quartz schists, they were traced along the strike. With one minor exception, no truncation of any of these horizons was found. Instead, they either pass beyond the limits of the map area or become impossible to follow because of metamorphic effects or lateral variations in sedimentary characteristics. The lack of any structural discordance is seen as evidence for the absence of the root zone of a major thrust fault in this area.

It is possible that the absence of truncated structures may not be sufficient proof against the existence of the Taconic fault in a given area. Recent work by Cady (personal communication) along the eastern edge of the klippe in the Rutland Valley has raised some question regarding the exact nature of the truncated structure of the Valley marbles. He found evidence indicating that the truncation is the result of Ordovician erosion of the calcareous series and unconformable overlap by Canajoharie shales of Ordovician age prior to thrusting. If this concept is correct, the sole of the thrust in certain localities separates slates of the two sequences which can be differentiated only by their fossil content; in other localities, notably the Brandon area, the thrust sheet appears to have ridden directly over the truncated structures of the Valley marbles. The detailed field work necessary to establish this relationship definitely has not been done.

It should be borne in mind, however, that the nature of the fault surface may be entirely different in the root zone from what it is in the area of the klippe. Marked discordances may prevail in the former zone, while the sole of the fault farther from the focus of dynamic activity may characteristically follow bedding surfaces without cross-cutting the structure to any significant degree.

It is quite true that the horizons which were followed represent a very small proportion of the entire stratigraphic column and that there is ample space left between them for all kinds of structural complexities.

Although the absence of truncated structures is purely negative evidence, it is believed that in conjunction with the other arguments this is a significant contribution to our understanding of the regional structure.

Consistent structural pattern.—A much more convincing point comes from a consideration of the consistent pattern of the structure over the whole map area. There is a notable parallelism of strike in the section extending from the top of the Ottauquechee formation west to the dolomite of the Pico Peak series as may be seen on the map. Such a homogeneous structural pattern would be anomalous in an area which contains a fault of the magnitude that has been postulated for the Taconic thrust. This point is illustrated in the Brandon area where the Taconic fault separates structurally distinct areas. It seems more likely that the quartz schists are made up of essentially conformable and parallel sediments which have not been complicated by major faulting.

The above points all lead to the conclusion that the roots of the Taconic fault do not exist east of the dolomite of the Pico Peak series. The following arguments support the thesis that a major structural discordance, probably the Taconic fault, occurs between this dolomite horizon and the Cheshire quartzite.

Absence of Cheshire quartzite on east side of range.—The heavily bedded Cheshire quartzite, from 300 to 800 feet thick, flanks the Valley marbles on the east for almost the entire length of the State. As determined from fossil evidence and from the structural relations at the Sucker Brook unconformity (Keith, 1932) it lies on the eastern limb of the Valley syncline. This must also be the western limb of an anticline which lies east of the mountain border. If the structure east of the Cheshire is a simple anticline, the Cheshire quartzite and the overlying marbles should reappear to the east on the eastern limb of the same anticline. Inasmuch as the Cheshire formation is persistent both in thickness and lithology for 200 miles along the strike, it is not likely that it pinches out or becomes generally unrecognizable in the short lateral distance represented by the arch of the anticline. Nevertheless, nothing comparable to the Cheshire formation has been found east of the range anywhere along the entire length of the Green Mountain sequence, with but one exception. In the North Adams area of northwestern Massachusetts Prindle and Knopf (1932, areal map) show the Green Mountain axis plunging south beneath the Valley sequence, while the Cheshire swings around the south end of the range in the form of the mirror-image of the letter J; for a short distance both the Cheshire quartzite and the overlying Rutland marbles appear on the east side of

the range before they are truncated by a fault structure, possibly the roots of the Taconic thrust.

To return to the Rutland area, the only mechanism, excepting non-deposition, which could account for the anomalous occurrence of a formation on only one flank of an anticline is a fault, which in some way does not permit that formation to appear at the surface on the other flank. In this case, the fact that the Cheshire formation is cut out for the entire length of the State implies a fault of considerable magnitude lying somewhere to the east.

Homoclinal structure of Green Mountains.—A similar conclusion comes from a consideration of the structure of the Green Mountain sequence. The evidence there indicates that the structure of the Green Mountain range in the latitude of Rutland is an eastward-dipping homocline which extends at least as far west as the dolomite of the Pico Peak series. The present investigation revealed no recurrence of this unique and persistent horizon which might be interpreted as part of the other limb of the fold. Furthermore, there is no positive evidence of a major fault in the Green Mountain sequence east of the outcrop of this horizon. If this analysis is correct, only the eastern flank of the complete anticlinal structure of the Green Mountains remains exposed in this locality. The simplest explanation for this anomalous relationship is that between the dolomite of the Pico Peak series and the Rutland valley a fault has cut off the western limb of the anticline.

Contrast of Green Mountain and Valley structures.—Further evidence for placing a major structural discordance between the Cheshire quartzite and the dolomite of the Pico Peak series can be seen by comparing the structural pattern of the Valley sequence with that of the Green Mountain sequence (Fig. 2). In the latitude of Rutland, the quartz schists apparently form a simple homocline, striking northwest and dipping east. The adjoining Marble Belt, as interpreted by Bain (Longwell, 1933, Pl. 13) has been intricately folded and cut by high-angle thrusts trending north-south. There is a conspicuous absence of parallelism between the trend of the Cheshire quartzite and that of the marker horizons which have been traced in the Green Mountain sequence.

From the above considerations, the present author concludes that the roots of the Taconic thrust fault most probably lie between the dolomite of the Pico Peak series and the Cheshire quartzite.

REVIEW OF CRITICAL AREAS

Critical field evidence in the present investigation is: (1) the trend of the dolomite of the Pico Peak series; (2) the trend of the quartzites at

the base of the Plymouth Union series; and (3) the nature of the contact of the Plymouth Union series and the Pinney Hollow schist. The first may be checked on the Sherburne Pass trail to Pico Peak and in the valley east of Blue Ridge Mountain; the basal quartzite of the Plymouth Union series may be traced south more or less continuously from a point three fourths of a mile east of the pass on the Mendon-North Sherburne road; the transition zone between the Pinney Hollow and the Plymouth Union series may be seen in the beds of brooks east of the Ottauquechee River north of Sherburne Center.

Foyles' (1938) recent work on Blue Ridge Mountain found no evidence for a major fault in that critical area. Metamorphic effects there have so obscured the original lithologic characteristics of the rocks that it is nearly impossible to trace out large-scale structures. A more promising area for further investigation is the terrane between Mt. Carmel and the Rutland valley. Not only are the rocks less metamorphosed there, but indications of several good calcareous marker horizons west of the dolomite of Mt. Carmel were found. It is strongly recommended that further intensive work be done in this area.

SUMMARY AND CONCLUSIONS

An area on the east side of the Green Mountain range between Plymouth and Rochester, Vermont, was studied in detail in the field in an effort to locate the roots of the Taconic thrust fault. While no evidence was found for the fault in that area, the writer believes that there is considerable evidence that the fault lies between the crest of the range and the Rutland valley. The arguments against the existence of the roots of the Taconic fault in the area studied are: (1) The contact of the Pinney Hollow schist and the Plymouth Union series, which Keith origially interpreted as the trace of the thrust plane, is apparently a simple, sedimentary transition; (2) no major fault truncation of any formations was observed within the area; (3) the structures of the Green Mountain sequence form a consistent regional pattern. If the thrust were east of the base of the sericite schists, the contrast in sedimentary facies of the Rutland calcareous sediments and the presumably contemporaneous argillaceous rocks east of the range is not explained. The fault therefore must lie west of the dolomite of the Pico Peak series. Positive arguments for placing the fault a short distance east of the Valley marbles are: (1) West of the Green Mountain range, the Cheshire quartzite and the immediately overlying Rutland marbles appear only on the western limb of an anticlinal structure; (2) in the center of the range, the formations of the quartz schists east of the dolomite of the Pico

Peak series appear only on the eastern limb of an anticline; (3) the structural pattern within the Green Mountain sequence is strikingly dissimilar to that in the Valley sequence. These features all indicate a major structural discordance between the dolomite of the Pico Peak series and the Cheshire quartzite. This discordance is most probably the roots of the Taconic thrust fault.

WORKS TO WHICH REFERENCE IS MADE

Adams, C. B. (1846) *Second annual report on the geology of the State of Vermont,* Burlington, Vt.

Cady, W. M. (1938) *Stratigraphy and structure of west-central Vermont* (Abstract), Geol. Soc. Am., Bull., vol. 49, p. 1870-1871.

Clark, T. H. (1934) *Structure and stratigraphy of southern Quebec,* Geol. Soc. Am., Bull., vol. 45, p. 1-20.

Cooke, H. C. (1937) *Thetford, Disraeli, and eastern half of Warwick map-areas, Quebec,* Canad. Geol. Survey, Mem. 211.

Dale, T. N. (1915) *The calcite marble and dolomite of eastern Vermont,* U. S. Geol. Survey, Bull. 589.

Fairbairn, H. W. (1932) *Structure and metamorphism of Brome County, Quebec,* Harvard University Doctor's Dissertation, p. 1-164.

Foye, W. G. (1919) *A report on the geological work within the Rochester, Vermont, quadrangle,* Vt. St. Geol., 11th Rept., p. 76-98.

Foyles, E. J. (1938) Unpublished investigation of Blue Ridge Mountain.

Hitchcock, E., *et al.* (1861) *Geology of Vermont,* Burlington, Vt., 2 vol. p. 1-988.

Kaiser, E. P. (1938) *Structural features of the northern part of the Taconic area,* (Abstract), Geol. Soc. Am., Bull., vol. 49, p. 1889-1890.

Kay, G. M. (1937) *Stratigraphy of the Trenton group,* Geol. Soc. Am., Bull., vol. 48, p. 233-302.

Keith, A. (1932) *Stratigraphy and structure of north-western Vermont,* Washington Acad. Sci., Jour., vol. 22, p. 357-379, 393-406.

Longwell, C. R., *et al.* (1933) *Eastern New York and western New England,* 16th Inter. Geol. Cong., Guidebook no. 1.

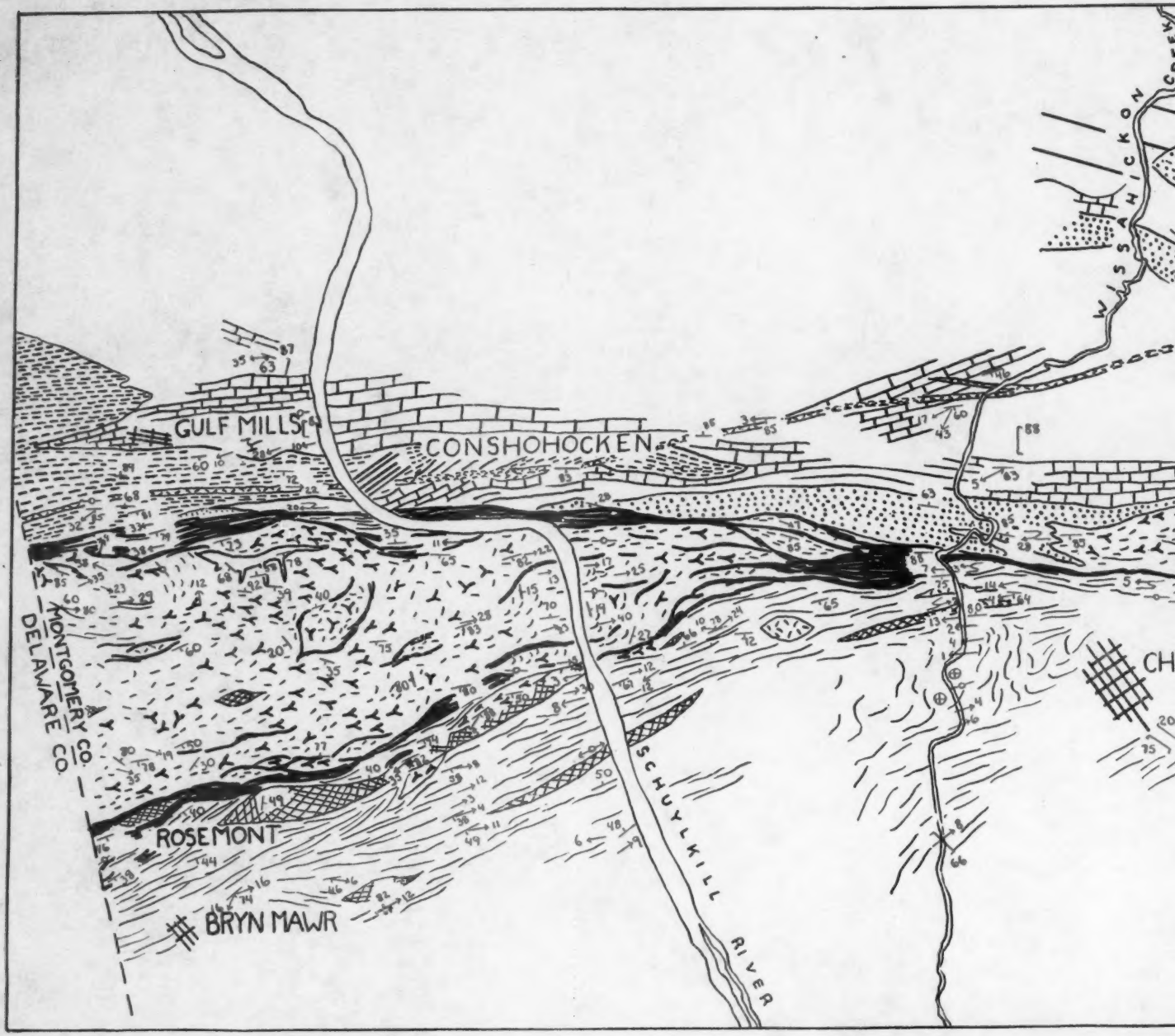
Perry, E. L. (1928) *The geology of Bridgewater and Plymouth townships, Vermont,* Vt. St. Geol., 16th Rept., p. 1-64.

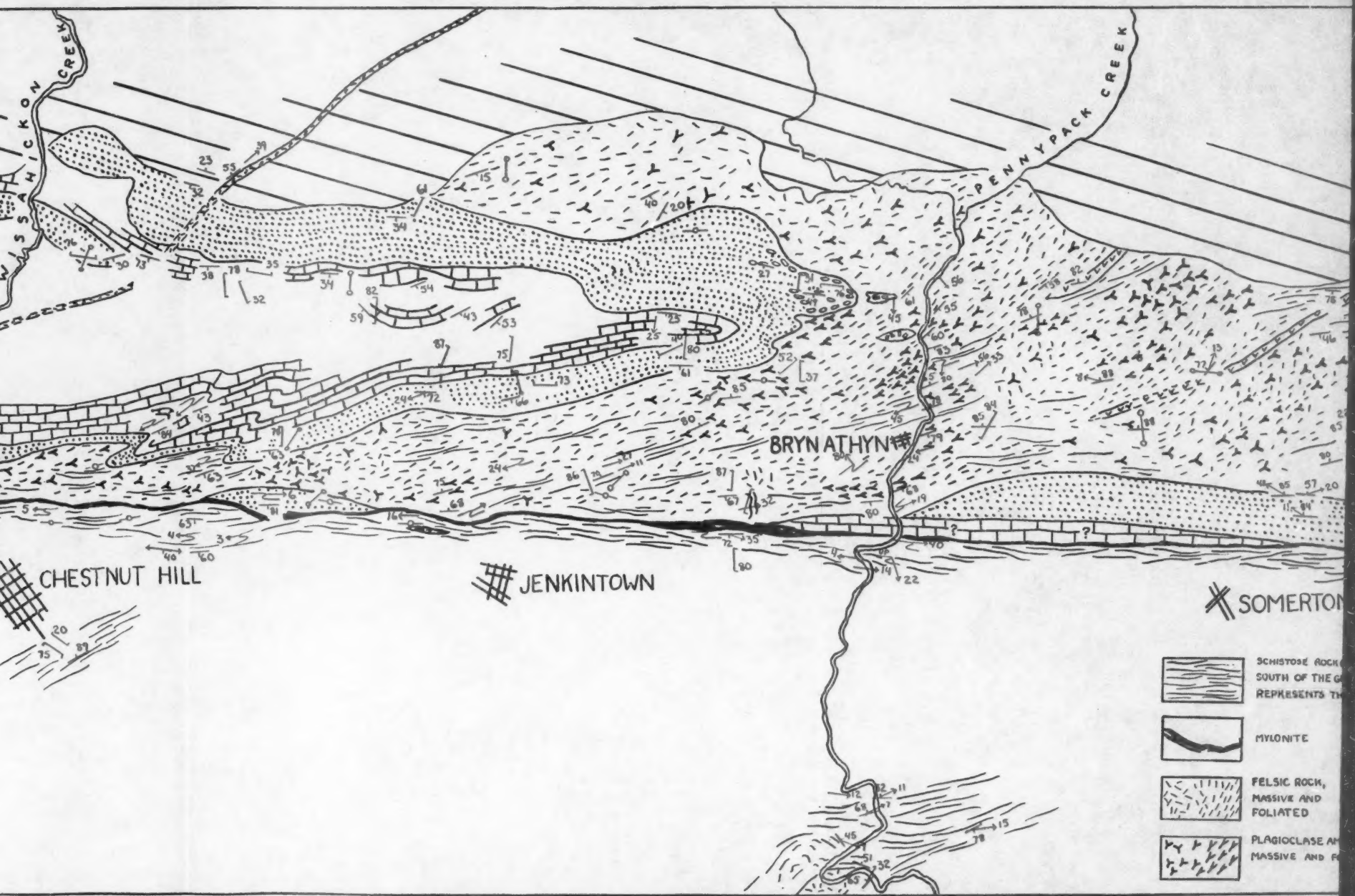
Prindle, L. M., and Knopf, E. B. (1932) *Geology of the Taconic quadrangle,* Am. Jour. Sci., 5th ser. vol. 24, p. 257-302.

Richardson, C. H., and Maynard, J. E. (1938) *Geology of Vernon, Guilford and Halifax townships, Vermont,* Vt. St. Geol., 21st Rept., p. 74-83.

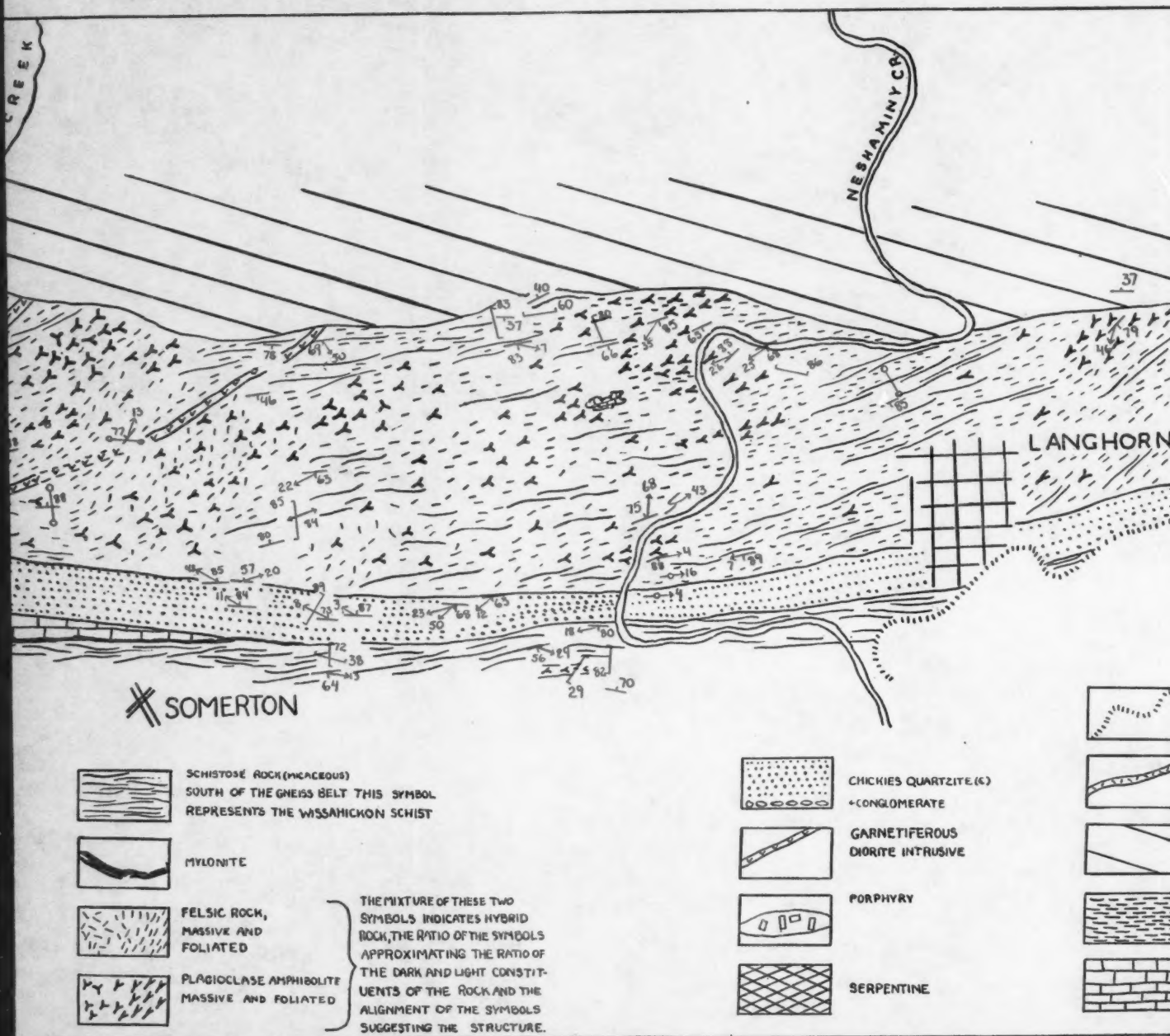
Whittle, C. L. (1894) *The occurrence of Algonkian rocks in Vermont and the evidence for their subdivision,* Jour. Geol., vol. 2, p. 396-429.

Massachusetts Institute of Technology, Cambridge, Mass.
Manuscript received by the Secretary of the Society, June 4, 1940.

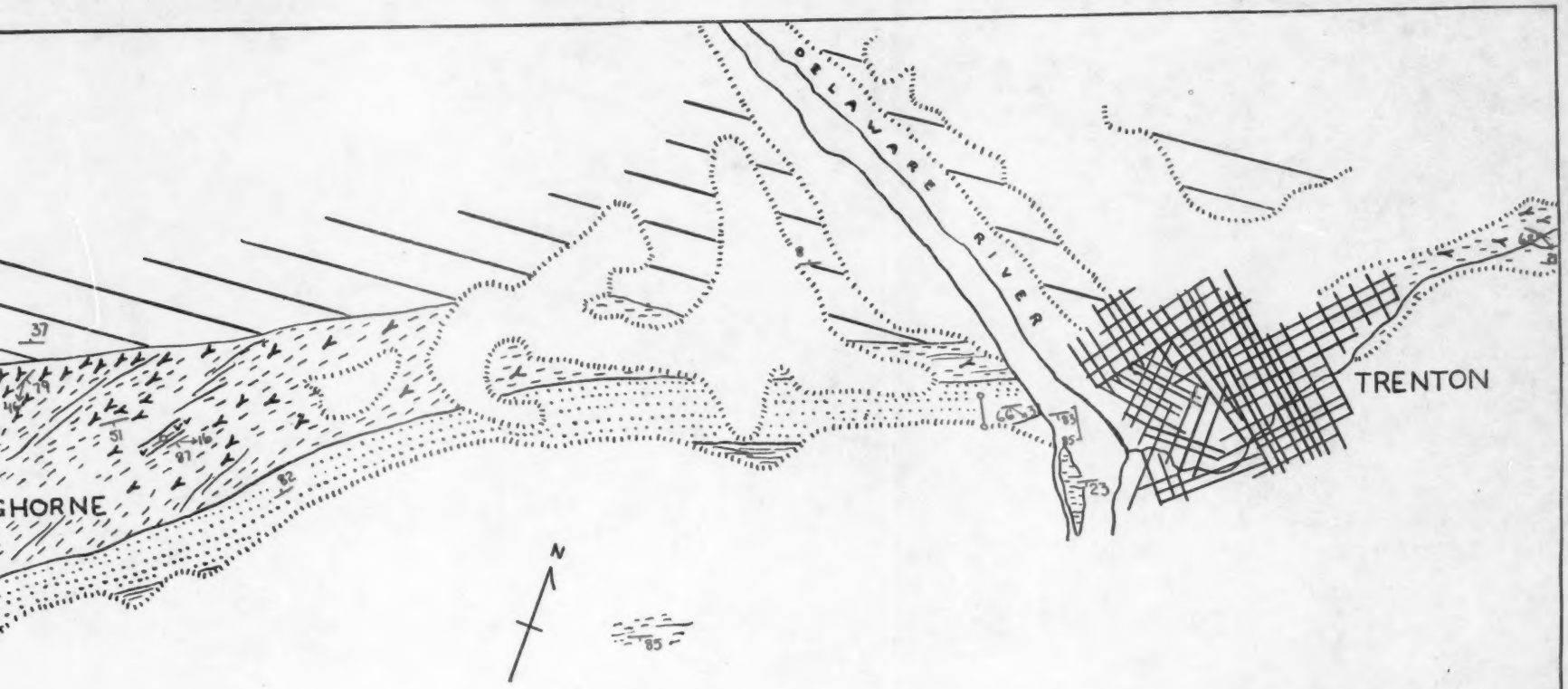

GULF MILLS
CONSHOHOCKEN
WISSAHICKON
MONTGOMERY CO.
DELAWARE CO.
ROSEMONT
BRYN MAWR
SCHUYLKILL RIVER

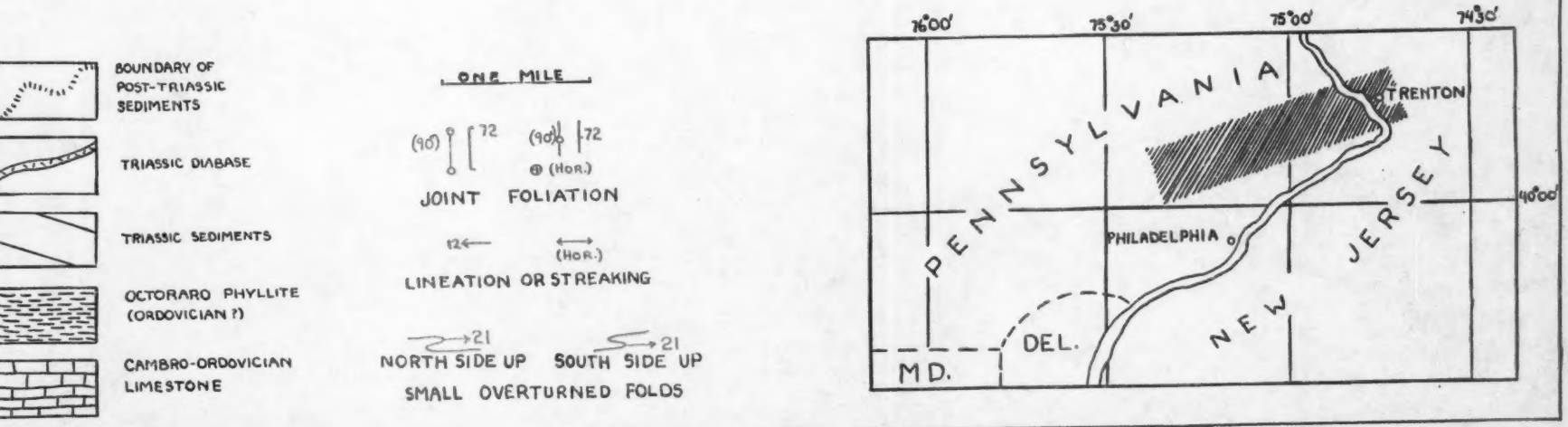


GEOLOGIC AND STRUCTURAL MAP OF GNEISS COMPLEX AND PARTS OF ADJACEN


NESHAMINY CR
LANGHORN
SOMERTON
SCHISTOSE ROCK (MICACEOUS) SOUTH OF THE GNEISS BELT THIS SYMBOL REPRESENTS THE WISSAHICKON SCHIST
MYLONITE
FELSIC ROCK, MASSIVE AND FOLIATED
PLAGIOCLASE AMPHIBOLITE MASSIVE AND FOLIATED
THE MIXTURE OF THESE TWO SYMBOLS INDICATES HYBRID ROCK, THE RATIO OF THE SYMBOLS APPROXIMATING THE RATIO OF THE DARK AND LIGHT CONSTITUENTS OF THE ROCK AND THE ALIGNMENT OF THE SYMBOLS SUGGESTING THE STRUCTURE.
CHICKIES QUARTZITE (C) +CONGLOMERATE
GARNETIFEROUS DIORITE INTRUSIVE
PORPHYRY
SERPENTINE

EX AND PARTS OF ADJACENT FORMATIONS NEAR PHILADELPHIA


DELAWARE RIVER
TRENTON
GHORNE
37
51
87
82
85
23


BOUNDARY OF POST-TRIASSIC SEDIMENTS
TRIASSIC DIABASE
TRIASSIC SEDIMENTS
OCTORARO PHYLLITE (ORDOVICIAN ?)
CAMBRO-ORDOVICIAN LIMESTONE
ONE MILE
(90) 72
JOINT
(90) 72
⊕ (HOR.)
FOLIATION
12
(HOR.)
LINEATION OR STREAKING
21
NORTH SIDE UP
21
SOUTH SIDE UP
SMALL OVERTURNED FOLDS

76°00'
75°30'
75°00'
74°30'
40°00'
PENNSYLVANIA
TRENTON
PHILADELPHIA
NEW JERSEY
DEL.
MD.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 667-694, 8 PL., 5 FIGS. MAY 1, 1941

MYLONIZATION OF HYBRID ROCKS NEAR PHILADELPHIA, PENNSYLVANIA

BY ELIZABETH ARMSTRONG

CONTENTS

	Page
Abstract	668
Acknowledgments	668
Location and boundaries of the region	669
Topography	669
Previous work in the region	669
Definitions	670
Petrographic	670
Structural	670
Localities cited	670
The hybrid rocks	673
Description and interpretation	673
Structure	676
Foliation	676
Axes of folds	676
Joints	676
The mylonites	677
Description and interpretation	677
Structure	680
Foliation	680
Axes of folds	680
Joints	681
Schistose zones east of Wissahickon Creek	681
Description and interpretation	681
Structure	682
Foliation	682
Axes of folds	682
Garnetiferous diorite intrusives	683
Quartz veins	683
Micaceous zones developed along boundary faults	684
Description and interpretation	684
Structure	685
Foliation	685
Axes of folds	685
Age relationships	686
Summary of conclusions	688
Works to which reference is made	688

ILLUSTRATIONS

Figure		Page
1.	Generalized diagram of mineral composition of the hybrid rocks	671
2.	Diagram of small overturned folds	672

Figure Page
3. Injection of amphibolite by felsic material south of Conshohocken 674
4. Distribution of poles of joint planes, plotted on a Schmidt net..... 677
5. Sketch of outcrop, a quarter of a mile southeast of Spring Mill, Pennsylvania.. 678

Plate Facing page
1. Geologic and structural map of the gneiss complex and parts of adjacent formations near Philadelphia..... 667
2. Outcrops of hybrid rock..... 674
3. Shear zones in less sheared rocks..... 675
4. Outcrops of mylonite..... 680
5. Photomicrographs..... 690
6. Photomicrographs..... 691
7. Photomicrographs..... 692
8. Photomicrographs of pseudotachylyte..... 693

ABSTRACT

The pre-Cambrian banded gneiss of a part of the Piedmont Plateau near Philadelphia is interpreted as a hybrid rock, resulting from the thorough injection of felsic material into a plagioclase amphibolite. The composition of the injecting solutions probably ranged from mostly water to that of an aqueous potash-silica-alumina magma. In the western part of the area vertical, winding mylonite zones cut these rocks. These zones outline lenticular masses of less sheared rocks and in some places attain a thickness of a quarter of a mile. East of Wissahickon Creek schistose shear zones run parallel to the nearly vertical foliation of the gneiss. A post-Ordovician high-angle fault forms the northern border of the gneiss west of Wissahickon Creek, crosses it and forms the southern border of the gneiss east of Wissahickon Creek. A branch of this fault bounds the western part of the gneiss on the south. The paper presents the results of a structural and petrographic study of a belt of the gneiss 37 miles long, extending from Trenton to Bryn Mawr.

ACKNOWLEDGMENTS

The field work upon which this paper is based was made possible by grants from the Department of Geology of Bryn Mawr College, and nearly 150 thin sections were made for this work by that department. The work was undertaken at the suggestion of Professor Edward H. Watson of Bryn Mawr College and has been carried on under his supervision. Professor Dorothy Wyckoff, also of Bryn Mawr, has helped the writer throughout the course of this work and especially in the preparation of this paper. The writer is indebted to Professor S. J. Shand of Columbia University for valuable discussions concerning the interpretation of some of the thin sections and for criticism of the manuscript. Others to whom the writer is indebted for critical reading of the manuscript are: Professor Emeritus Florence Bascom of Bryn Mawr College, Professor Ida H. Ogilvie and Miss Delia W. Marble of Barnard College, and Dr. Eugene Cameron of Columbia University. Professor Ernst Cloos of the Johns Hopkins University read the paper preparatory

to its publication, and the writer is especially indebted to him for his very careful and detailed criticism.

LOCATION AND BOUNDARIES OF THE REGION

The area under discussion lies in Montgomery and Bucks counties in southeastern Pennsylvania and in Burlington County in New Jersey and is part of the Piedmont Plateau. Mapped on the Norristown, Germantown, and Burlington quadrangles, it is bounded on the southwest by the Montgomery County-Delaware County line and extends northeastward to 4 miles northeast of the Delaware River at Trenton.

The hybrid rocks constitute a belt varying in width from less than a quarter of a mile to about 2¾ miles and are bounded on the southeast by the Wissahickon schist and on the northwest by the Chickies quartzite, the Cambro-Ordovician limestones, the Octoraro phyllite, and the Triassic sediments (Bascom, 1909a; 1909b).

TOPOGRAPHY

The gneiss forms a ridge running approximately N. 65° E. and is cut by the Schuylkill River, Wissahickon Creek, Pennypack Creek, and Neshaminy Creek which flow across it in steep-sided valleys. Where the ridge is narrow and well defined, it is known as Buck Ridge. Most of the small streams run parallel to it, many of them following either the border faults or the mylonite zones within the gneiss.

PREVIOUS WORK IN THE REGION

Early mention of the gneiss was made in H. D. Rogers' *Geology of Pennsylvania* (1858) where it is called a syenite. In this work a portion of the mylonites of the present paper was interpreted as basal Paleozoic sediments. In the Second Geologic Survey of Pennsylvania (1881) C. E. Hall refers to the gneiss as "syenite and gneissic granites or granitic gneisses".

In 1885 H. Carvill Lewis interpreted the gneiss "to be of pure eruptive origin, consisting of syenites, acid gabbros, trap granulites and other igneous rocks, often highly metamorphosed". Lewis recognized the mylonites as such, describing them as "identical with the mylonite of Lapworth", whose paper proposing the term (Lapworth, 1885) had just then been published.

A geologic map of the region appeared in 1909 in the geologic folios for Philadelphia (Bascom, 1909b) and for Trenton (Bascom, 1909a), in which the gneiss is described and correlated with the pre-Cambrian Baltimore gneiss of Maryland.

DEFINITIONS

PETROGRAPHIC

FELSIC MATERIAL: All the intruded or injected material whose invasion into the plagioclase amphibolite produced the hybrid gneisses. The composition of this material probably ranged from mostly water to an aqueous potash-silica-alumina magma.

FELSIC ROCK: Rock consisting predominantly of felsic material, in all cases including altered plagioclase or biotite or hornblende derived from the plagioclase amphibolite. (*See* mineral analyses 5-8 of Table 1.)

PLAGIOCLASE AMPHIBOLITE: Rock composed essentially of hornblende and andesine with as much as 5 per cent pyroxene in some places. Small amounts of quartz and accessories are present throughout. (*See* mineral analyses 1-4 of Table 1.)

The hybrid rocks show every gradation in composition from the plagioclase amphibolite through intermediate types to the felsic rock (Fig. 1).

TABLE 1.—*Mineral composition of the hybrid types by volume per cent*

Analyses were made by linear measurements across thin sections with an integrating stage. Small amounts of potash feldspar, biotite, and hornblende are included with the accessories in some analyses.

No.	Rock type	Locality	Plagioclase	Potash feldspar	Quartz	Biotite	Hornblende	Accessories
1.	Plagioclase amphibolite	6	12	..	16	..	65	7
2.		5	38	..	2	1	50	9
3.		3	32	..	11	..	46	11
4.		4	34	..	11	11	36	8
5.	Felsic rock	7	33	31	15	15	..	6
6.		4	30	29	28	11	..	2
7.		7	17	35	31	12	..	5
8.		7	15	29	33	20	..	3
9.	Pegmatite	5	3	58	38	..	..	1

STRUCTURAL

SMALL OVERTURNED FOLDS: Small, repeated overturned folds, with no observed relation to any larger folds, occur throughout the region, especially in the schistose shear zones and the mylonite zones. They are interpreted as indicating differential movement as illustrated in Figure 2. The map symbol used for such folds is also illustrated in Figure 2.

LOCALITIES CITED

1. Arrowmink Creek, 1/16 mi. south of its passage beneath Conshohocken State Road
2. East side of Pennypack Creek, quarry ¼ mi. north of Paper Mills railroad station
3. On road, east side of Schuylkill River, ¾ mi. northwest of railroad station at Miquon (formerly Lafayette)
4. On road, west side of Schuylkill River, 2 mi. northwest of Roseglen railroad station
5. On road, east side of Schuylkill River, 1 mi. northwest of Miquon (Lafayette) railroad station

6. Philadelphia and Western railroad (not shown on topographic map) cut 1⅛ mi. S. 13° W. of Gulf Mills
7. On road, east side of Schuylkill River ⅝ mi. northwest of Miquon (Lafayette) railroad station

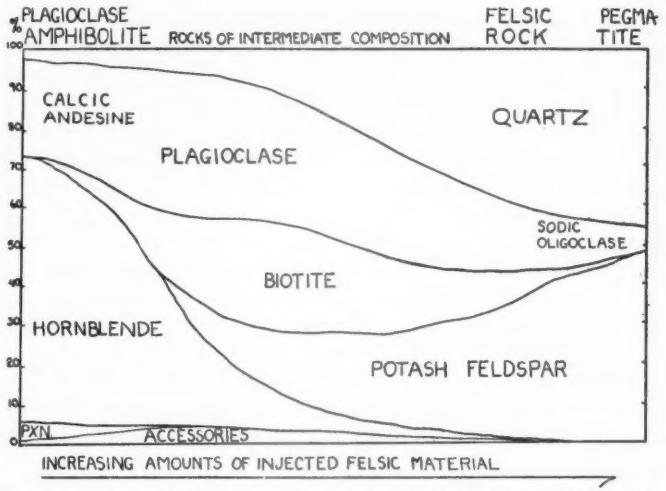


FIGURE 1.—*Generalized diagram of mineral composition of the hybrid rocks*

8. East side of Pennypack Creek, 500 feet south of Paper Mills railroad station, outcrop on creek and on railroad
9. East side of Pennypack Creek at Paper Mills railroad station
10. Quarry at Bryn Athyn (formerly Alnwick) west side of Pennypack Creek
11. Junction of stream and road, ⅜ mi. S. 42° W. of Bryn Athyn (Alnwick) railroad station
12. ⅒ mi. south of Bryn Athyn (Alnwick) railroad station
13. ⅝ mi. west of Paper Mills railroad station
14. ⅒ mi. south of a point on the Schuylkill River ¾ mi. downstream from the bridge at Conshohocken
15. Quarry in Chickies quartzite, ¾ mi. N. 38° E. of Laverock railroad station
16. 1¼ mi. S. 65° W. of the railroad station at Miquon (Lafayette)
17. Yard of St. Thomas's parochial school, Lancaster Ave., Rosemont
18. River Road and Arrowmink Creek, south side of the Schuylkill River opposite Conshohocken
19. De Haven Ave., 1000 feet south of Matson's Ford Road 1 mi. S. 61° E. of Gulf Mills
20. Railroad cut, west side of the Schuylkill River, ⅜ mi. N. 81° W. of the railroad station at Miquon (Lafayette)
21. Quarry 1/16 mi. N. 45° W. of Lenni Mills railroad station, about 5 miles northwest of Chester

22. On road, west side of the Schuylkill River, ¼ mi. north of the bridge at Conshohocken
23. Montgomery Ave., 100 feet southeast of Ithan Ave.
24. River road, west side of the Schuylkill River, 1/10 mi. north of Young's Ford Road

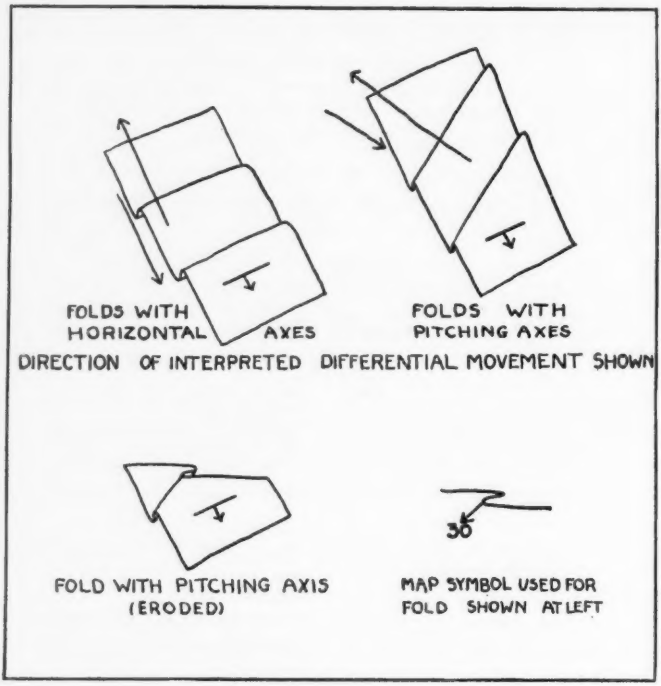

FOLDS WITH HORIZONTAL AXES
FOLDS WITH PITCHING AXES
DIRECTION OF INTERPRETED DIFFERENTIAL MOVEMENT SHOWN
FOLD WITH PITCHING AXIS (ERODED)
30
MAP SYMBOL USED FOR FOLD SHOWN AT LEFT

FIGURE 2.—*Diagram of small overturned folds*

25. 1⅛ mi. S. 50° E. of Gulf Mills
26. Railroad cut 1/10 mi. northwest of Laverock railroad station
27. Railroad cut at Woodbourne railroad station
28. Road cut ¾ mi. S. 82° W. of Woodbourne railroad station
29. Railroad cut 1¾ mi. N. 35° E. of the cross roads at Siles (west of Langhorne)
30. Quarry on Assanpink Creek 3 mi. N. 53° E. of the north bridge at Trenton (Route 1, by-pass)
31. Quarry ⅛ mi. north of Mill Creek at Holland P. O. (west of Langhorne)
32. Quarry on golf course, northwest corner of Langhorne
33. West bank of Delaware River, ¼ mi. south of north bridge (Route 1, by-pass) at Trenton

34. Road cut ⅜ mi. N. 50° W. of Meadow Brook railroad station (southwest of Bryn Athyn)

35. Van Artsdalen's quarry ⅞ mi. due north of the crossroads at Siles (west of Langhorne)

THE HYBRID ROCKS

DESCRIPTION AND INTERPRETATION

Abundant field and microscopic evidence indicates that injection of felsic material into a plagioclase amphibolite (Harker, 1932, p. 311) has produced the banded and massive types of the gneiss. The character of the injection varies from magmatic intrusion of bodies with well-defined contacts to hydrothermal penetration. In some of the mixed rocks felsic and mafic bands alternate (Pl. 2, fig. 1); elsewhere the injected material forms a network of dikelets in the plagioclase amphibolite (Pl. 2, figs. 2, 3), and over much of the region the mixing has been so complete that a nearly homogeneous rock has been formed. All gradations of composition occur in each mixed type. Probably both the condition of the wall rock at the time of injection and the composition of the felsic material injecting it were factors determining the character of the resulting mixed rock. The megascopically homogeneous types probably resulted where the plagioclase amphibolite was hot or the injecting material was aqueous and full of volatile constituents that facilitated permeation of the wall rock; the coarsely inhomogeneous hybrid rocks probably resulted where the wall rock was cold or the injecting magma relatively dry. Figure 3 is a sketch of an outcrop that shows injection phenomena of the latter type.

Where least affected by the injection, the plagioclase amphibolite is composed chiefly of hornblende, andesine (ca. An_{35}), and quartz with smaller amounts of biotite and accessories (Anal. 1-4, Table 1). Accessories are magnetite, epidote, zircon, pyrite, and, in many cases, abundant euhedral sphene, white by reflected light. This rock will be referred to as the plagioclase amphibolite, the amphibolite, or the mafic rock. It varies from granitoid to schistose.

This rock contains as much as 5 per cent hypersthene in some places between Neshaminy Creek and Trenton near the north border of the gneiss (Loc. 27-30). In these rocks hypersthene and hornblende have crystallized simultaneously as shown by hypersthene crystallized around euhedral hornblende and hornblende crystallized around euhedral hypersthene (Pl. 5, fig. 1). Throughout the region altered grains of pyroxene occur sparsely in the plagioclase amphibolite. In some cases the pyroxene remnant is surrounded by antigorite with a rim of chlorite (Pl. 5, fig. 5); in other cases the pyroxene has been replaced by hornblende. This alteration of the pyroxene has been observed in rocks in which the

plagioclase is unaltered and a negligible amount of quartz and potash feldspar is present, which suggests that the alteration was not caused by the later injection of felsic material.

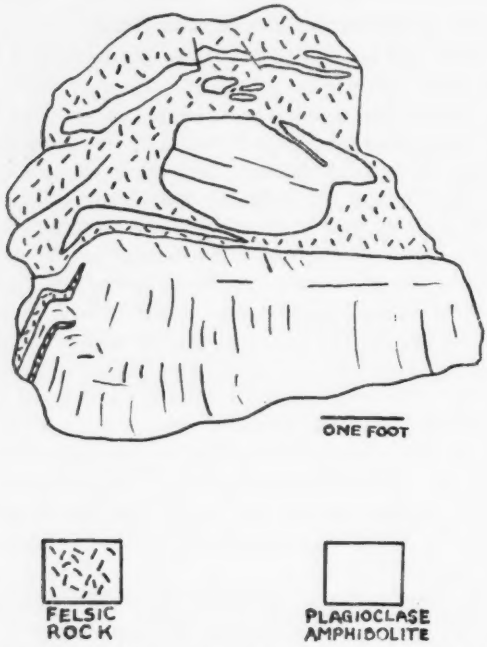


FIGURE 3.—*Injection of amphibolite by felsic material south of Conshohocken*

These mafic rocks are probably of igneous origin. The field evidence does not indicate that the amphibolite was formed by alteration of a limestone in the manner described by Adams and Barlow (1910), although in some places it contains as much as 10 per cent carbonate. Metamorphosed limestone crops out in a single small quarry, known as Van Artsdalen's quarry, a mile west of Neshaminy Creek (Loc. 35). Crystals of amphibole have been formed in this limestone by injection of felsic material, but these are isolated from the masses of plagioclase amphibolite whose contacts with the limestone show chilled borders. Next to these contacts the texture of the limestone is coarsened. These phenomena indicate magmatic intrusion rather than hydrothermal replacement.

FIGURE 1. PEGMATITE INJECTION IN BANDED HYBRID ROCK
(Locality 5)

FIGURE 2. CONTORTED BANDED HYBRID ROCK
(Locality 5)

FIGURE 3. MAFIC ROCK INJECTED BY FELSIC MATERIAL
On Conshohocken State Road, ½ mi. south of Conshohocken.

OUTCROPS OF HYBRID ROCK

FIGURE 1. STEEP-DIPPING FRACTURES IN FOLIATED FELSIC ROCK
(Locality 34)

FIGURE 2. CONTACT BETWEEN MYLONITE AND LESS SHEARED ROCK
(Locality 1)

SHEAR ZONES IN LESS SHEARED ROCKS

Where felsic material has penetrated the plagioclase amphibolite, as at the localities illustrated in Plate 2, rocks of intermediate composition have developed. A summary of the observable reactions follows: biotite has replaced hornblende; plagioclase has become more sodic with formation of epidote in enclosed laths a few hundredths of a millimeter long (Pl. 5, fig. 2); potash feldspar has replaced plagioclase (Pl. 5, fig. 2); and quartz, potash feldspar, and plagioclase have replaced hornblende and biotite.

The alteration of hornblende to biotite and the change in composition of the plagioclase are closely associated with the amount of felsic material in the hybrid rock. Where the proportion of quartz and feldspar is high the alteration of hornblende to biotite is well advanced, the plagioclase is sodic oligoclase (ca. An_{22}), and potash feldspar is present. Where the proportion of quartz and feldspar is low, only small amounts of biotite have been formed, the plagioclase is more calcic (ca. An_{30}), and little or no potash feldspar occurs.

Even among those felsic rocks that look like granite gneisses in the field, none is without schlieren composed of epidote-bearing oligoclase, biotite, and accessories, and in many cases such schlieren also contain hornblende. The surrounding rock is made up of quartz, potash feldspar, and oligoclase. The potash feldspar forms embayments in the oligoclase (Pl. 5, fig. 2), and some larger potash feldspar grains have small, irregularly bounded cores of oligoclase enclosing epidote. Analyses 5-8 of Table 1 show the composition of these rocks.

Both in the amphibolite and in the more felsic rocks, quartz-hornblende replacement aggregates occur whose shape is that of a pre-existing, replaced mineral, probably pyroxene (Pl. 5, fig. 6). Primary hornblende may occur in the same rock (Pl. 5, fig. 6), showing that the rock originally contained both pyroxene and hornblende, as it still does in some places (Pl. 5, fig. 1).

East of Wissahickon Creek garnet is abundant in the hybrid rocks of intermediate composition at a few localities. These garnetiferous rocks are of two types: either they occur at the border of a predominantly amphibolitic mass, as at Paper Mills (Loc. 2) where some of the rock contains more than 50 per cent garnet; or they occur in a gneiss composed of reddish biotite, blue quartz, oligoclase, and garnet (Loc. 31, 32). Rock of the same texture, mineral composition, and appearance as this second garnetiferous type crops out on the Delaware River (Loc. 33) within the area mapped as Wissahickon schist in the Philadelphia Folio (Bascom, 1909b). West of Wissahickon Creek the rocks of intermediate composition predominate, and several areas of the plagioclase amphibolite occur, whereas, east of Wissahickon Creek, the felsic types predominate.

Pegmatite dikes cut the hybrid rocks at some places (Pl. 2, fig. 1). Their contacts may be sharp or gradational, and a dike that cuts the banding for part of its length may change its direction and become one of the light bands of the banded rock. The composition of a specimen from the dike figured in Figure 1 of Plate 2 is given as analysis 9 of Table 1.

STRUCTURE

Foliation.—The foliation of the hybrid rocks is due to the parallel orientation of hornblende grains in the amphibolite and of schlieren in the felsic rocks. Foliated intermediate types show a well-defined banding parallel to the foliation. West of Wissahickon Creek these bands are folded. For the most part, the folds are open with gently dipping limbs, but some of them are overturned toward the northwest (Pl. 2, fig. 1), and some are complexly contorted (Pl. 2, fig. 2).

East of Wissahickon Creek, however, steep dips predominate, and the average strike is approximately N. 60° E. It is not everywhere possible to determine whether the foliation is primary or secondary. In some places it is certainly primary. Near Bryn Athyn the steep foliation of the amphibolite parallels that of the felsic rocks, and dark xenoliths in the felsic rocks are also parallel to this structure. Here the structure in the felsic rock must be a flow foliation whose structure was controlled by that of the foliated amphibolite into which it was injected.

Gently dipping foliation was observed a mile northeast of Jenkintown (Loc. 34) in felsic gneiss, cut by steep joints striking approximately N. 60° E. (Pl. 3, fig. 1). These joints have developed along secondary foliation planes which occur at intervals of a few inches. The foliation is due to recrystallization, especially of biotite, resulting from slight displacement along these planes. In view of this evidence it seems possible that in other parts of the area the steep foliation now observable may have obliterated a former, gently dipping foliation.

Axes of folds.—In the banded hybrid rocks the axes of folds strike N. 50°–70° E. and with few exceptions pitch 10°–30° NE. This structure is common to all the folds, whether large or small, open or recumbent.

Joints.—The orientation of joints in the gneiss between Chestnut Hill and Conshohocken is shown in Figure 4(A). A similar diagram of joints in the Cambro-Ordovician limestones north of the gneiss is included for comparison. In addition to joints parallel to the foliation, which are not plotted, the most prominent joint system throughout the gneiss, represented by the maximum at 1 in Figure 4, strikes north and dips vertically or nearly so. The joint system indicated by the concentration of points at 3 in Figure 4 is parallel to the mylonite zones in this

part of the region and probably originated during a late phase of the same deformation that produced the mylonites. The concentration of points at 2 in Figure 4 represents a system of oblique joints. The symbol X marks the average orientation of the fold axes.

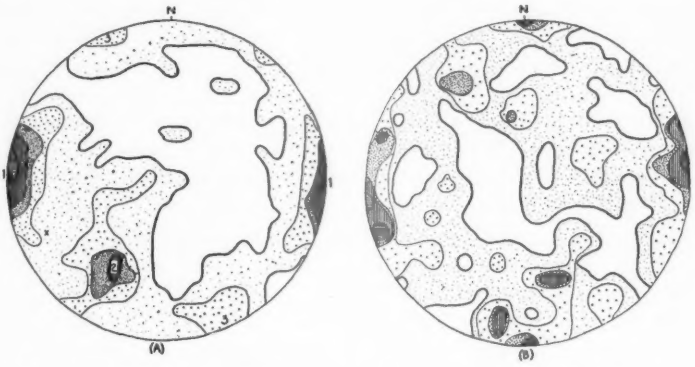


FIGURE 4.—*Distribution of poles of joint planes, plotted on a Schmidt net*

(A) Joints in the gneiss, from 841 measurements between the meridians 75° 12′ 30″ W. and 75° 17′ 30″ W.

(B) Joints in the Cambro-Ordovician limestones, from 170 measurements.

THE MYLONITES

DESCRIPTION AND INTERPRETATION

Narrow, winding, nearly vertical, mylonite zones cut the hybrid rocks of the western part of the area.[1] They strike N. 50°–70° E. and outline lenticular masses of less sheared, massive, or banded rock. Repetition of this structure occurs on all scales of magnitude. Lenticles as long as a mile may be seen on the map (Pl. 1), lenticles a few inches to 2 feet in length may be seen in outcrops (Fig. 5; Pl. 4, figs. 1, 2), and lenticles a fraction of a millimeter to several millimeters long may be seen under the microscope (Pls. 5, 6, 7).

This repetition of the lens shape in all sizes has been cited by Harker (1932, p. 167) as characteristic of "discontinuous deformation."

The thickness of the mylonite zones varies from a quarter of a mile (Pl. 1) to a fraction of a millimeter (Pl. 6, fig. 3). Their structure ranges from thinly laminated to massive,[2] and their color from white through light grays and reddish browns to dark brownish gray. Most common

[1] Waters and Campbell (1935) give a glossary of terms used for different types of mylonites and an extensive bibliography of papers concerning crushed rocks.

[2] "Cataclasite" of Grubenmann and Niggli (1924).

FIGURE 5.—*Sketch of outcrop, a quarter of a mile southeast of Spring Mill, Pennsylvania*

Nearly vertical mylonite zones, in part isoclinally folded, with lenticles of massive and banded (here folded) hybrid rock.

are light grayish-brown mylonites that break into slabs about half an inch thick and contain small, lenticular, yellow-white feldspar grains.

They all have a dull luster which is due to their extremely fine texture. All but the most intensely crushed [3] (Pl. 7, fig. 2) contain feldspar (Pl. 6, figs. 5, 6; Pl. 7, figs. 1, 3) or hornblende (Pl. 6, fig. 4) or both in len-

[3] "Ultramylonite" of Staub (1915) and Quensel (1916).

ticular or pod-shaped grains, most of which are porphyroclasts (grains that were not completely pulverized). During deformation plagioclase grains bend, apparently with differential movement between the twinning lamellae (Pl. 5, fig. 3), but the more brittle quartz breaks. The lenticular shape must be assumed by a grain if it is to survive mylonization.[4] It is a "streamlined" shape past which the surrounding crystal mush moves with a minimum of disturbance. As a result the mylonites look porphyritic, since lenticular grains remain relatively large and the others are pulverized.

A few grains are porphyroblasts, crystallized subsequent to the crushing of the rock, as indicated by their ungranulated borders, unstrained crystal structure, and lack of conformity with the fluidal structure of the mylonite. Epidote porphyroblasts are common in the mylonites (Pl. 6, figs. 5, 6; Pl. 7, fig. 4); some bands are composed almost exclusively of finely crushed biotite and larger, subhedral epidote grains (Pl. 7, fig. 2). Apatite and sphene are less common as porphyroblasts than is epidote.

Recrystallization of hornblende and feldspar is relatively rare. One hornblende crystal 3 centimeters long was found in banded mylonite (Loc. 1), and a few large, fresh feldspar crystals occur in the compact, massive type of mylonite.

Quartz grains are very rare either as porphyroclasts or porphyroblasts. Quartz was one of the first constituents of the rock to become granulated by crushing and recrystallized as elongated grains in thin, straight-sided bands (Pl. 5, fig. 4; Pl. 7, fig. 4), never as thick lenticles or euhedral grains.

The transition zone from less sheared rock to mylonite is less than 5 feet in most places (Pl. 3, fig. 2) and may be a fraction of a millimeter (Pl. 6, figs. 2, 3). Masses of less sheared rock are separated from each other by winding sheer zones in some of which the shearing is only moderate (Pl. 6, fig. 1), in others more extreme, but not yet such as to form the most compact type of mylonite (Pl. 5, fig. 3; Pl. 6, fig. 4); in still others true mylonization has taken place (Pl. 6, figs. 5, 6).

Examination of one thin section containing essential potash feldspar, quartz, oligoclase, and biotite (Anal. 7, Table 1) in which the transition from mortar structure to mylonite occurs within less than a millimeter (Pl. 6, fig. 2) disclosed the following changes: by bending, recrystallization, and some granulation the biotite became well oriented, with feldspar (both oligoclase and potash feldspar) in elongated grains between the biotite plates. With increasing granulation, first the biotite became

[4] Both "mylonization" and "mylonitization" appear in the literature. "Mylonization", which has been used by Clough, Maufe, and Bailey (1909, p. 630) and by Jehu and Craig (1925, p. 635) is preferred by the writer because it is shorter. Its meaning is as clear as that of the other form.

triturated, then the quartz, and finally the feldspar. The mixing of these pulverized constituents produces a pale-brown *purée parfaite* (Termier and Boussac, 1911) with a uniform pleochroism that indicates that the biotite is well oriented.

The commonest type of mylonite has a groundmass composed chiefly of biotite as described above, although white mylonites composed entirely of quartz and feldspar occur, as well as dark-gray mylonites composed of hornblende, plagioclase, and quartz with a good deal of biotite. The biotite mylonite appears to be most readily formed from the hybrid rocks which contain as an essential constituent the biotite which is nearly lacking in the plagioclase amphibolite and the felsic rock (Fig. 1). Many of the less sheared lenticles, as shown on Plate 1, are either predominantly of the felsic rock without the inclusion of much of the amphibolite constituents or predominantly of the plagioclase amphibolite without much felsic injection. The intermediate biotite-rich rock in which the mylonite zones are generally formed lies between these two, both areally and in composition.

The presence of biotite is believed to be one of the causes, rather than exclusively the result, of mylonization because microscopic examination shows that the fine biotite of the mylonites is formed by the crushing of larger biotite grains and because the mylonite zones are most commonly associated with the biotitic intermediate hybrid rocks.

STRUCTURE

Foliation.—The strike of the foliation of the mylonites varies because they form winding zones, the average being about N. 65° E. The dip, with few exceptions, is nearly vertical.

The laminated mylonites show folding varying from occasional small overturned folds (Pl. 4, fig. 2) to repeated isoclinal folds (Fig. 5). The amplitude of the folds is usually less than a foot, and it seems possible that they could have been produced at the same time as the mylonization. An oil film on water moving slowly along a confined path, as in a gutter, illustrates the development of laminae by differential movement and the folding of those laminae by the same movement. Sander (1930, p. 69) describes related phenomena.

Axes of folds.—The axes of small, overturned folds in the mylonites, as in the less sheared gneiss, strike N. 50°–70° E., pitching northeastward at low angles. On the north and south borders of the gneiss in the western part of the region, however, fold axes in the mylonites pitch southwestward. The significance of this will be discussed under the heading Age Relationships.

FIGURE 1. LENTICULAR MASS OF LESS SHEARED HYBRID ROCK IN MYLONITE
(Locality 14)

FIGURE 2. MYLONITE SHOWING STRUCTURES INDICATING
DIRECTION OF MOVEMENT
(Locality 14)

FIGURE 3. MYLONITE SHOWING HORIZONTAL DRAG-FOLD AXES
On Wissahickon Creek, a mile northwest of Chestnut Hill.

Joints.—Steep joints parallel and normal to the foliation predominate in the mylonites, and at many localities diagonal sets of joints occur which break the mylonite into rhomb-shaped slabs (Pl. 4, fig. 3).

SCHISTOSE ZONES EAST OF WISSAHICKON CREEK

DESCRIPTION AND INTERPRETATION

Mylonites are rare east of Wissahickon Creek. The few occurrences discovered are less than a foot thick and cannot be traced for more than a few feet along the strike. Closely spaced schistose partings, however, parallel the foliation of the gneissoid felsic rock which underlies most of the eastern portion. These bands of mica schist might be interpreted as shear zones in the felsic igneous rock or as septa of a schist into which the felsic material was injected.

The gneissoid rock here is composed of sodic oligoclase (ca. An_{12}) and quartz with varying amounts of potash feldspar. Many of the oligoclase grains show cores containing abundant small epidote laths and rims devoid of epidote. Differences in composition of the two parts of the feldspar were not observed. The quartz and some of the feldspar grains are granulated, and most of the feldspar grains are broken.

The schistose bands, most of which are less than a quarter of an inch thick, contain about equal amounts of muscovite and biotite, both in long, thin plates. The bands are winding and include lenticles of feldspar and quartz around which the mica flakes are wrapped. Steep shear zones occur in this part of the area (Pl. 3, fig. 1), and it seems more probable that these schistose bands represent shear zones in the felsic rock than that they are included septa of an earlier schist. Further pertinent discussion is given under the heading Age Relationships.

At Paper Mills and Bryn Athyn there are bands of predominantly schistose rock several feet wide (Loc. 9, 10, 11, 12). Furthermore, near Paper Mills layers a foot or two thick that look like typical gray quartzite crop out (Loc. 13). At many localities in the eastern part of the region graphite gneiss may be found. These rocks are of particular interest because, in the field, they look more like metamorphosed sediments than does any other type of the gneiss in this region. The graphite gneiss is composed of oligoclase and quartz in about equal proportions. The diameter of the quartz grains varies from ½ to 3 millimeters, and strain shadows are common. The oligoclase is interstitial and extensively sericitized. Biotite and garnet are the accessories. Some of the garnet grains are embayed by quartz. The graphite occurs along fractures and between the grains, indicating that it was introduced after the rock was formed. An outcrop of pegmatite which cuts the graphite gneiss shows a concentration

of graphite at its contact with the finer-textured gneiss. Whether the graphite was introduced by the pegmatite (Smith, 1906) or was only redistributed by hydrothermal solutions (Smith, 1906) is not known.

The quartzitic rock is composed of quartz grains varying in diameter from 0.1 to 1 millimeter. Some of them are nearly round, but most of them are extremely irregular. All the grains show strain shadows, and there is evidence that the smaller grains have originated by granulation of the larger ones. Sericite fills the interstices between the quartz grains, but none of the plagioclase from which it was presumably derived is left. Crushed garnet is abundant, the fragments occurring in elongated groups with interstitial chlorite.

On the basis of both field and microscopic evidence, these two rock types could be interpreted either as hydrothermally altered sediments or as fine-textured, felsic igneous rocks. The question of their origin must be left open until further evidence is discovered.

The possibility has been mentioned that the narrow schistose bands east of Wissahickon Creek resulted from shearing. If this is so, why did shearing produce mylonites in the western part of the area and mica schists in the eastern part? The two parts of the area are separated by a fault whose amount of displacement is unknown. The rocks south of the fault (*i.e.*, west of Wissahickon Creek) have moved down with respect to the north side as indicated by small overturned folds in many places in and near the fault zone (Pl. 1). The western part of the gneiss, then, may have been closer to the surface than the eastern part when the mylonites were formed. If the schistose zones were formed by shearing at the time of the mylonization, they are the deep-seated equivalents of the mylonites. An alternative possibility is that these schistose shear zones east of Wissahickon Creek are contemporaneous with those described under the heading Micaceous Zones Developed Along Boundary Faults. Because these schists are confined to the boundary faults in the western part of the area, however, and because the winding structure of the thin schistose partings is closely comparable to that of the mylonites, it seems probable that the schistose zones east of Wissahickon Creek were formed at the time of the mylonization. Renewed movement along these zones at the time of the boundary faulting is believed to have taken place.

STRUCTURE

Foliation.—The schistose zones east of Wissahickon Creek are parallel to the foliation of the felsic gneiss.

Axes of folds.—The axes of most of the small overturned folds in these zones pitch southwestward at low angles. A few exceptions near Neshaminy Creek are plotted on Plate 1.

GARNETIFEROUS DIORITE INTRUSIVES

The hybrid rocks are cut by hornblende-quartz-labradorite dikes containing a little pyroxene and a great deal of garnet. The texture of these dikes is in part subophitic. One dike, a mile southwest of Gulf Mills, is about one eighth of a mile thick at its widest section and can be traced for nearly a mile along its strike which is parallel to the foliation of the gneiss. Smaller dikes extend southeastward from the main one for less than half a mile, following the steep cross joints in the banded gneiss which strike about N. 50° W. in this part of the region. Two other large dikes and many small dikelets occur in the eastern part of the area (Pl. 1).

Both hypersthene and a pyroxene that is probably diopside occur in these rocks and are in part replaced by hornblende or by a fibrous aggregate of uralite(?). In the narrower dikes the plagioclase is andesine (ca. An_{44}). A few small dikes cutting the banded gneiss half a mile north of Rosemont (Loc. 23) belong to this group of intrusives. A garnetiferous diabase composed essentially of labradorite (ca. An_{53}) and hypersthene crops out near the southern border of the gneiss on the west side of the Schuylkill River. Its foliation is nearly normal to that of the gneiss, and its pyroxene content is much higher than that of the intrusives just described. Whether it is related to these intrusives or to the hypersthene-bearing types described with the hybrid rocks is not known.

A tough, compact, porphyritic rock underlies a region a quarter of a square mile in area west of Pennypack Creek at Paper Mills (Loc. 13). Prism-shaped, fibrous aggregates of sericite half a centimeter long occur in a fibrous groundmass composed predominantly of hornblende. No outcrop of the porphyry was found, and its relationship to the surrounding rocks is unknown.

QUARTZ VEINS

Quartz veins cut both the mylonite and the garnetiferous dikes at localities less than 5 miles apart. In the mylonites they conform to the fluidal structure of the crushed rock, and although many of the quartz grains show strain shadows they are not crushed. In the garnetiferous dikes the quartz veins are rimmed by dark, hornblende-rich zones which may be as much as three-quarters of an inch wide. In some places this is greater than the width of the quartz vein itself.

Quartz veins also occur in the Cambrian Chickies quartzite to the north, filling cross joints that are presumably tension joints formed at the time of the Appalachian folding (Loc. 15).

The relative ages of these three sets of quartz veins are unknown.

MICACEOUS ZONES DEVELOPED ALONG BOUNDARY FAULTS

DESCRIPTION AND INTERPRETATION

A fault, described by Bascom (1909b) as the Cream Valley fault, bounds the gneiss on the north in the western half of the area, crosses it just east of Wissahickon Creek, and continues on the south side of the gneiss in the eastern part of the region where it is called the Huntingdon Valley fault (Bascom, 1909b). A second fault bounds the gneiss on the south in the western part of the region and joins the Huntingdon Valley fault just east of Wissahickon Creek. In this paper this southern fault will be called the Rosemont fault, from the town of Rosemont which is situated on the fault line. It is a branch of the Cream Valley-Huntingdon Valley fault. Both are steep faults. Small overturned folds along the Huntingdon Valley fault and along the Rosemont fault indicate movement downward on the south. Those along the Cream Valley fault show movement in both directions, probably indicating that there has been more than one period of slipping along this fault and that the downthrow side has not always been the same. The Cream Valley fault is mapped (Bascom, 1909b) as bifurcating again at the western boundary of the region.

Along the Rosemont fault (Loc. 16, 17) a mylonite with muscovite-spangled foliation planes occurs. These plates of muscovite are 1 to 5 millimeters long and have been deformed subsequent to crystallization. The chief constituents of this spangled mylonite are quartz and feldspar. Since the muscovite occurs in large, only slightly deformed plates whereas the quartz and feldspar are granulated, the muscovite must have formed after mylonization (Pl. 7, fig. 5). A similar rock in which the muscovite plates are somewhat more deformed (Pl. 7, fig. 6) crops out where the Cream Valley fault crosses Wissahickon Creek. Here a few grains of blue-green tourmaline occur in small prisms. These rocks are not predominantly schistose and are not as much recrystallized as Lapworth's augen schist (1885) or Quensel's mylonite gneiss (1916).

In the mylonites in the gneiss mass, muscovite was not observed in hand specimen and it appears only rarely in thin sections. The occurrence of the muscovite mylonites along the Rosemont and Cream Valley faults suggests that the development of the muscovite is associated with the faulting, possibly resulting from the rise of hydrothermal solutions along the fault planes.

Along the Cream Valley fault south and southwest of Conshohocken a schistose zone 20 to 30 feet thick crops out (Loc. 18, 19). South of Conshohocken (Loc. 18) it is bordered on either side by a narrow band of flinty crush rock containing veins of pseudotachylyte (Shand, 1916) a fraction of a millimeter thick (Pl. 8). The schist here is a muscovite-

chlorite schist containing garnet and blue-green tourmaline. The micas were deformed after crystallization.

Since muscovite, chlorite, and tourmaline are found in the Octoraro phyllite (Pl. 1) and muscovite and garnet, with some chlorite, in the Wissahickon schist, this schistose zone south of Conshohocken may be a lenticle of either of these rocks caught in the fault zone and sheared. Other corroborative evidence is the occurrence of folded bands of minute inclusions in several sodic oligoclase grains in this rock.

An alternative interpretation is that the mica of this schistose zone was developed in the same way as that of the muscovite mylonites. The structural data which follow are pertinent to this problem of interpretation.

STRUCTURE

Foliation.—The foliation of the muscovite schists along the Cream Valley fault and of the muscovite mylonites strikes N. 50°–70° E. and dips steeply toward the south in most places.

Axes of folds.—The axes of small overturned folds in most of these schistose rocks pitch southwestward at low angles, mostly less than 20 degrees. In the Wissahickon schist, for a few hundred feet south of the Rosemont fault and south of the Cream Valley fault, the axes pitch southwestward. Farther south in the Wissahickon schist, however, they pitch northeastward. Briefly, the axes of small overturned folds pitch northeastward in the gneiss and in the schist except in the vicinity of the faults where the pitch is to the southwest.

The southwestward-pitching folds were apparently formed at the time of the faulting, obliterating the previously formed northeastward-pitching folds which characterize those parts of the gneiss and schist not affected by the faulting. In one outcrop a few hundred feet south of the Rosemont fault (Loc. 20) small overturned folds with both northeastward- and southwestward-pitching axes occur. In the quartzitic layers the northeastward-pitching fold axes are well marked while the southwestward-pitching ones are faint, whereas the opposite is true in the schistose layers. Here the later deformation, forming the southwestward-pitching folds, strongly affected the less competent, schistose layers although it did not much affect the more competent, quartzitic layers. The quartzitic layers retained their earlier eastward-pitching folds, which, in the weaker schistose layers, became almost obliterated.

The schistose zones in the gneiss which have been compared to the mylonite zones (Schistose Zones East of Wissahickon Creek) show westward-pitching fold axes. For this reason it is believed that there was renewed movement along them at the time of the boundary faulting.

Ten miles south of this region, where the strike of the Wissahickon schist is north, the small overturned folds in the quartzitic layers of the Wissahickon schist have nearly vertical axes, while the schistose layers show folds pitching northward at low angles (Adolph Meier, oral communication). This seems to suggest an earlier deformation which formed steep-pitching fold axes and a later one, affecting chiefly the incompetent layers, producing gently pitching fold axes.

AGE RELATIONSHIPS

Metamorphosed limestone crops out in a single small quarry known as Van Artsdalen's quarry 2½ miles west of Langhorne. The area underlain by limestone is probably less than one acre, and no other limestone was found within the gneiss belt in this region. This rock has been correlated (Bascom, 1909a) with the Franklin limestone. The features observed in Van Artsdalen's quarry, which have been described, under The Hybrid Rocks, show that the plagioclase amphibolite intrudes the metamorphosed limestone. With this one exception, the rock into which the original mafic magma was intruded is unknown.

From the field evidence it would seem possible that some of the schistose shear zones east of Wissahickon Creek were included septa of an earlier schist. The only schist in this region is the Wissahickon schist which borders the gneiss to the south in many places (Pl. 1). In this schist plagioclase amphibolite intrusive sheets occur, injected by felsic material. The phenomena of hybridization in these rocks (Postel, 1940, p. 135) are similar to those described earlier in this paper. Excellent exposures of hybrid rock may be seen along the west bank of the Delaware River at Trenton (Loc. 33) for more than a quarter of a mile south of the north boundary of the Wissahickon schist (Bascom, 1909a). Felsic rock predominates, but a band of amphibolite with felsic injection producing abundant epidote occurs a short distance south of the Route 1 by-pass bridge at Trenton and is visible from the bridge. Very little schistose rock crops out at this locality. The heavy-mineral assemblage in the schistose zones in the gneiss indicates, however, that these zones are probably not septa of the Wissahickon schist. The most diagnostic heavy mineral of the Wissahickon schist, a pinkish brown tourmaline, was absent in four specimens of the schistose gneiss from different localities, whereas three heavy minerals characteristic of the gneiss—namely, epidote, pink pleochroic zircon, and sphene—were present.[5]

[5] The heavy-mineral analyses of these rocks were made for the writer as part of a general heavy-mineral survey of the middle Atlantic States being carried out by Lincoln Dryden and Clarissa Dryden of Bryn Mawr College.

The injection of the felsic material in the gneiss belt was accompanied by plastic deformation of the mixed rocks with folding overturned toward the northwest (Pl. 2, figs. 1, 2). This was followed by cataclastic deformation along the nearly vertical mylonite zones and shearing with recrystallization along the schistose zones.

The mylonite zones have been deformed repeatedly. The period of major mylonization followed the major injection of felsic material, as is shown by the many occurrences of microcline porphyroclasts and pulverized quartz as well as abundant field evidence. Minor hydrothermal injection subsequent to mylonization is indicated by the recrystallization of hornblende to well-developed biotite along some shear zones and the presence of ungranulated quartz veins in the mylonite. Strain shadows in the quartz of these veins resulted from renewal of deformation as did the slight granulation of some of the post-mylonization epidote porphyroblasts.

The Cambrian Chickies quartzite occurs in a southwestward-pitching syncline, floored by the gneiss. The gneiss must therefore be pre-Cambrian. The only exposed contact between the gneiss and the Chickies quartzite is at Neshaminy Creek (Pl. 1) where the two are separated by a steep shear zone about 15 feet thick.

Whether the mylonization preceded the deposition of the Paleozoic sediments is not known, but the structural evidence presented below is pertinent to this problem.

In the Cambrian and Ordovician sediments the axes of small overturned folds are parallel to the axes of the larger folds, pitching 15°–30° SW. These folds are post-Ordovician, probably formed during the Appalachian revolution. In the hybrid rocks, however, the axes of large and small folds pitch 15°–30° NE., a pitch direction very rarely found in the Paleozoic sediments.

If the northeastward-pitching folds are pre-Cambrian in this region and the southwestward-pitching folds post-Ordovician, then these structures may be useful locally in dating periods of deformation.

The fold axes in the mylonites pitch northeastward. In the schistose zones east of Wissahickon Creek both northeastward- and southwestward-pitching fold axes occur. The fold axes in the Wissahickon schist pitch northeastward except near the boundary of the gneiss, where the pitch is to the southwest, especially in the more micaceous layers. The boundary fault zones of the gneiss have southwestward-pitching fold axes. They cut out Cambro-Ordovician sediments and must be late or post-Ordovician.

The garnetiferous dikes are probably of post-mylonization age. The largest one in the western part of the area occurs near the edge of the

thickest mylonite zone but does not show intense shearing. This evidence is not conclusive, since parts of the hybrid rock that are near mylonite zones are also relatively unsheared (Pl. 3, fig. 2).

SUMMARY OF CONCLUSIONS

The belt of gneiss extending from Trenton southwestward to the Delaware County boundary line is composed of hybrid rocks formed by the intrusion of felsic material into a plagioclase amphibolite of igneous origin. The rock into which the mafic magma was intruded is not known. These hybrid rocks were compressed into folds overturned toward the northwest (Pl. 2, figs. 1, 2), probably while the felsic material was being injected. The formation of mylonites along numerous high-angle faults occurred west of Wissahickon Creek after the folding of the hybrid rocks. The many schistose zones east of Wissahickon Creek may have been formed by shearing at the same time.

The garnetiferous diorite intrusives that cut the hybrid rocks probably came in after the mylonization and before the deposition of the Paleozoic sediments, but the evidence concerning their age is still too meager for any definite decision to be made. Quartz vein injection followed their intrusion.

The Appalachian revolution again subjected the region to compression, producing folds that are overturned toward the northwest in the Paleozoic sediments, but, unlike the similarly overturned folds of the pre-Cambrian hybrid rocks, pitch southwestward.

After the Appalachian folding the Rosemont and Cream Valley-Huntingdon Valley faults were formed. This was, then, a second episode of high-angle faulting following compression. The development of abundant muscovite and westward-pitching fold axes distinguish these fault zones from the mylonite zones.

WORKS TO WHICH REFERENCE IS MADE

Adams, F. D., and Barlow, A. E. (1910) *Geology of the Haliburton and Bancroft areas, Province of Ontario,* Geol. Survey Canada, Mem. 6.

Bascom, F. (1909a) U. S. Geol. Survey, Geol. Atlas, Trenton Folio (no. 167).

——— (1909b) U. S. Geol. Survey, Geol. Atlas, Philadelphia Folio (no. 162).

Clough, C. T., Maufe, H. B., and Bailey, E. B. (1909) *The cauldron-subsidence of Glen Coe,* Geol. Soc. London, Quart. Jour., vol. 65, p. 611-678.

Grubenmann, V., and Niggli, P. (1924) *Die Gesteinsmetamorphose,* Gebrüder Borntraeger.

Hall, C. E. (1881) *Second Geologic Survey of Pennsylvania,* Harrisburg.

Harker, Alfred (1932) *Metamorphism,* Methuen, London.

Jehu, T. J., and Craig, R. M. (1925) *Geology of the Outer Hebrides, Part II,* Royal Soc. Edinburgh, Tr., vol. 53, Part III (no. 30), p. 615-641.

Lapworth, C. (1885) *The Highland controversy in British geology: its causes, course, and consequences,* Nature, vol. 32, p. 558-559.

Lewis, H. Carvill (1885) *Some examples of pressure-fluxion in Pennsylvania,* Nature, vol. 32, p. 559-560.

Postel, A. W. (1940) *Hydrothermal emplacement of granodiorite near Philadelphia,* Acad. Nat. Sci., Philadelphia, Pr., vol. 92, p. 123-152.

Quensel, Percy (1916) *Zur Kenntnis der Mylonit-bildung, erläutert an Material aus dem Kebnekaisegebiet,* Geol. Inst. Upsala, Bull., vol. 15, p. 91-116.

Rogers, H. D. (1868) *Geology of Pennsylvania,* D. Van Nostrand, New York, p. 74-76.

Sander, Bruno (1930) *Gefügekunde der Gesteine,* p. 67-69, Springer, Vienna.

Shand, S. J. (1916) *The pseudotachylyte of Parijs (Orange Free State),* Geol. Soc., London, Quart. Jour., vol. 72, p. 198-221.

Smith, G. O. (1906) *Graphite in Maine,* U. S. Geol. Survey, Bull. 285, p. 480-483.

Staub, R. (1915) *Petrographische Untersuchungen im westlichen Berninagebirge,* Vierteljahrsschrift d. Naturf. Gesell. im Zurich, vol. 60, p. 55-336, (especially p. 71-91).

Termier, P., and Boussac, J. (1911) *Sur les mylonites de la région de savone,* Compt. Rend., Ac. des Sci., vol. 152, p. 1550-1556.

Waters, A. C., and Campbell, C. D. (1935) *Mylonites from the San Andreas Fault zone,* Am. Jour. Sci., 5th ser., vol. 29, p. 473-503.

Barnard College, New York, N. Y.

Manuscript received by the Secretary of the Society, November 22, 1940.

EXPLANATION OF PLATES 5-8

PLATE 5.—PHOTOMICROGRAPHS

Figure 1.—Plagioclase amphibolite: hypersthene, hornblende, andesine. (× 40).
Figure 2.—Microcline replacing oligoclase containing epidote. (× nicols, × 40).
Figure 3.—Bent plagioclase grain in sheared hybrid. (× nicols, × 80).
Figure 4.—Mylonite with plagioclase porphyroclasts and recrystallized quartz bands. (× nicols, × 80).
Figure 5.—Pyroxene altered to antigorite and chlorite. (× 40).
Figure 6.—Granular aggregate of hornblende and quartz. (× 40).

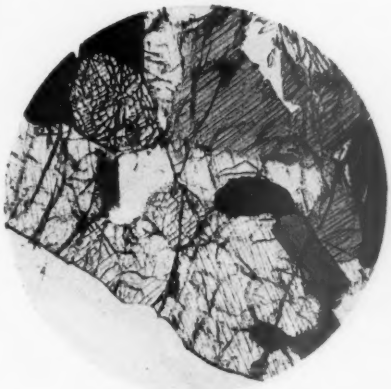


FIGURE 1

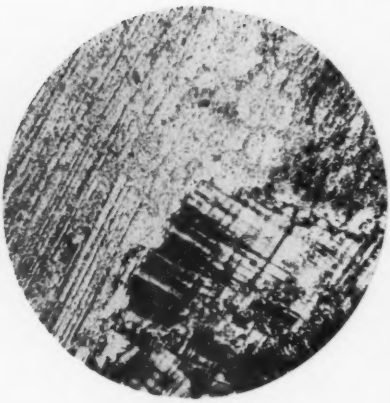


FIGURE 2

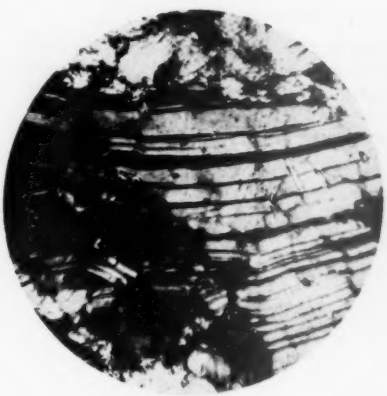


FIGURE 3

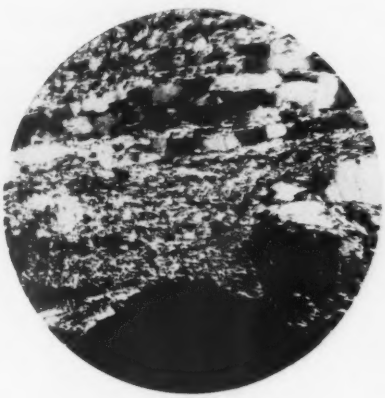


FIGURE 4

FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS

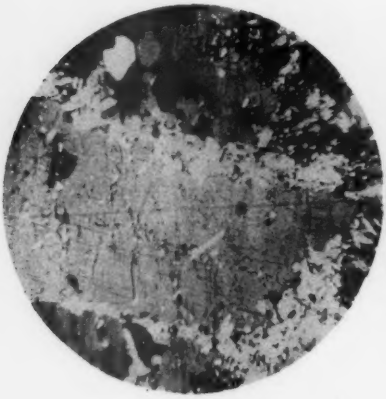


FIGURE 1

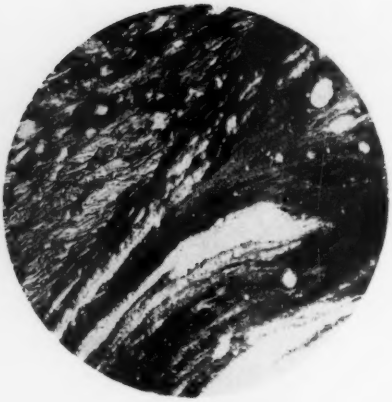


FIGURE 2

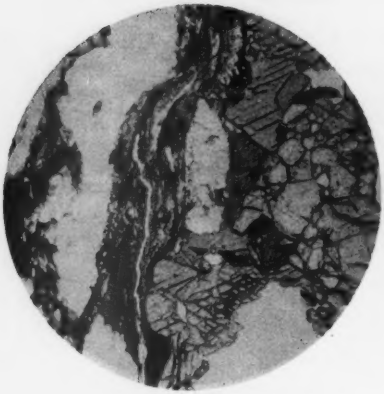


FIGURE 3

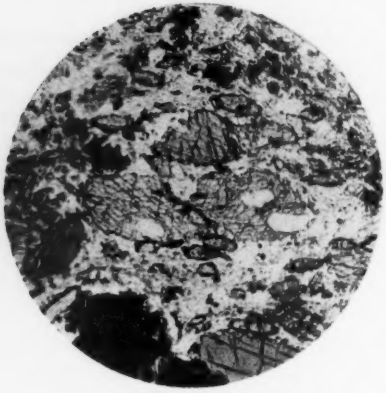


FIGURE 4

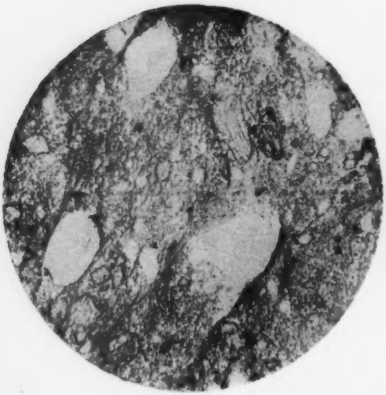


FIGURE 5

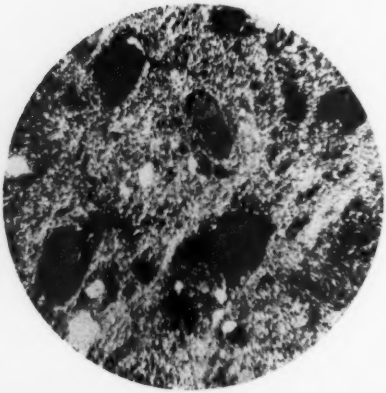


FIGURE 6

PHOTOMICROGRAPHS

PLATE 6.—PHOTOMICROGRAPHS (× 40)

Figure 1.—Granulation of borders of grains. (× nicols).
Figure 2.—Mylonite zone in sheared hybrid rock.
Figure 3.—Narrow mylonite zone in hybrid rock.
Figure 4.—Hornblende porphyroclasts in mylonite.
Figure 5.—Feldspar porphyroclasts and epidote porphyroblast in mylonite.
Figure 6.—Same as figure 5, × nicols.

PLATE 7.—PHOTOMICROGRAPHS

Figure 1.—Feldspar grains with magnetite "tails." (× 28).
Figure 2.—Ultramylonite with epidote porphyroblasts. (× 40).
Figure 3.—Feldspar porphyroclasts in mylonite with epidote porphyroblast. (× 40).
Figure 4.—Epidote porphyroblast and recrystallized quartz bands in mylonite. (× 80).
Figure 5.—Muscovite wrapping feldspar porphyroclast. (× 40).
Figure 6.—Deformed muscovite in mylonite. (× nicols, × 40).

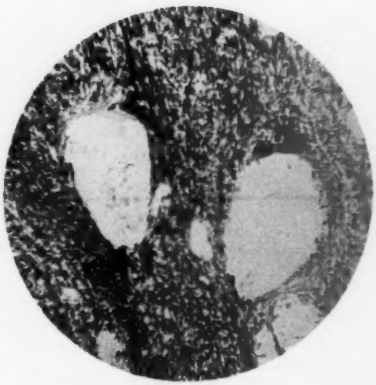


FIGURE 1

FIGURE 2

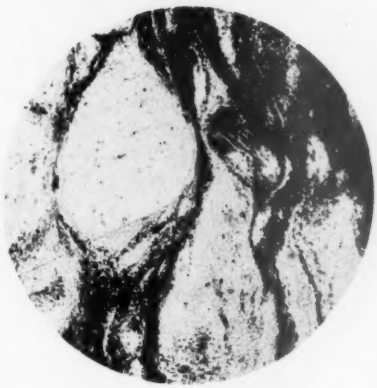


FIGURE 3

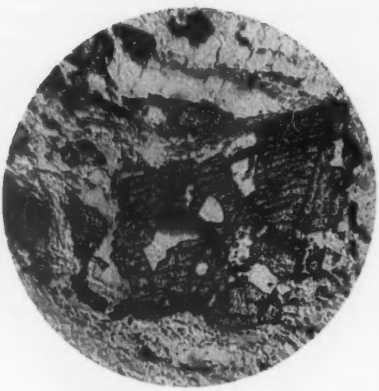


FIGURE 4

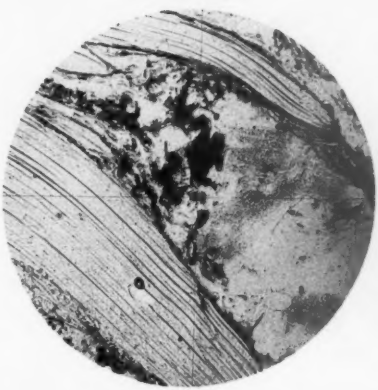


FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS

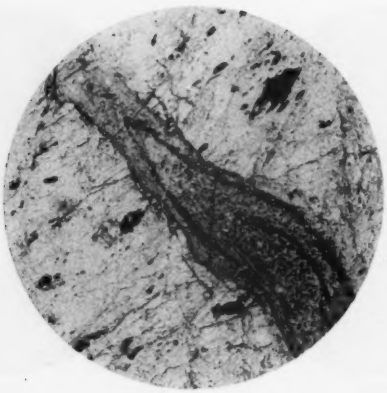


FIGURE 1

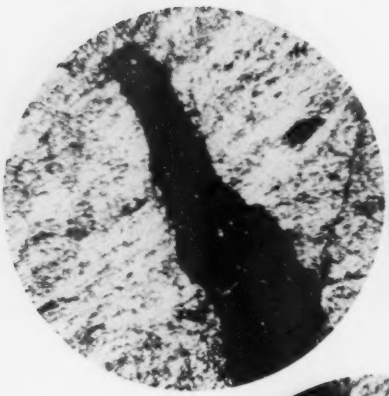


FIGURE 2

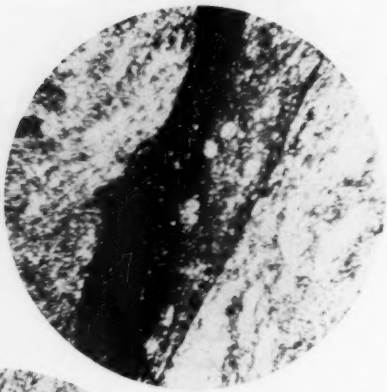


FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS OF PSEUDOTACHYLYTE

PLATE 8.—PHOTOMICROGRAPHS OF PSEUDOTACHYLYTE (× 30)

Figures 2, 3, and 4 illustrate adjoining or nearly adjoining parts of the same thin section. Figure 1 is of the same area as Figure 2, showing flow structures not observable with crossed nicols.

Figure 3.—Transition from mylonite to pseudotachylyte, the pseudotachylyte form parallel to the foliation of the mylonite.

Figure 4.—Pseudotachylyte branching from the main band to cut across the foliation of the mylonite.

Figure 2.—Blunt end of this branch.

Figures 2, 3, and 4 are between × nicols.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 695-720, 4 PLS., 1 FIG. MAY 1, 1941

GEOLOGY OF THE STONY MOUNTAIN STOCK, SAN JUAN MOUNTAINS, COLORADO

BY MC CLELLAND DINGS

CONTENTS

	Page
Abstract	696
Introduction	696
Topography	697
General geologic setting	697
Pre-Cambrian quartzite	699
Bedded volcanic series	700
General statement	700
San Juan tuff (Miocene)	700
Silverton volcanic series (Miocene)	700
Potosi volcanic series (Miocene)	701
Governor diorite	701
General statement	701
Rock description	702
Chemical and modal analyses	704
Hornblende monzonite facies	705
General statement	705
Rock description	705
Chemical and modal analyses	705
Theory of origin	707
Gabbro	707
General statement	707
Rock description	708
Modal analyses	709
Chemical analysis	710
Quartz monzonite facies	711
Origin of quartz monzonite	711
Stony Mountain diorite	712
General statement	712
Rock description	712
Modal analyses	713
Rhyolite	714
General statement	714
Rock description	714
Age of rhyolite	715
Dikes	715
General statement	715
"Radial" andesite dikes	715
Aplite dikes	717
Clastic dikes	717

Page

"Non-radial" andesite porphyry, andesite, and dacite dikes 718
Rhyolite dikes 718
Summary 719
Works to which reference is made 719

ILLUSTRATIONS

Figure Page

1. Geologic map of Stony Mountain region 698

Plate Facing page

1. Stony Mountain and quartzite inclusions 700
2. Exposures of gabbro, diorite, and rhyolite 701
3. Contacts 714
4. Photomicrographs 715

ABSTRACT

Volcanic rocks of Miocene age belonging to the San Juan, Silverton, and Potosi series have been intruded by a stock which shows a sequence of intrusions from diorite to gabbro to diorite, and finally to rhyolite. The stock is excellently exposed for a vertical distance of 1900 feet. The first diorite intrusion was locally transformed after emplacement to a hornblende monzonite facies, probably by gaseous transfer. Many radiating andesitic dikes are related to the diorite. Gabbro, the next intrusion, has an associated quartz monzonite facies that was differentiated at depth. Many fragments and large masses of pre-Cambrian quartzite were carried up in the gabbroic magma. Clastic dikes unquestionably formed from material derived from lower-lying conglomeratic beds penetrate the gabbro. A uniform, fine-grained diorite with associated andesite and dacite dikes represents the third intrusion. The last intrusion is a rhyolite which probably was emplaced after a considerably greater time interval than existed between the diorites and gabbro. Two rocks from the stock were chemically analyzed, and numerous modal analyses are tabulated. The undisturbed, nearly horizontal attitude of surrounding volcanic rocks has interest in regard to the mode of emplacement of the stock.

INTRODUCTION

During the late spring and summer of 1936 the writer studied the Stony Mountain stock, located about 5 miles southwest of Ouray in the San Juan Mountains of southwestern Colorado.

Since it lies within a region of considerable importance for gold and silver mining, the stock has been frequently mentioned in publications on economic geology, but its rock types and igneous history have not been studied in detail. Cross (1899) mapped the general outline and briefly described the stock, but in the absence of detailed work the main body was mapped as a gabbro-diorite of Tertiary age. He believed an associated rhyolite to be Algonkian, but the present writer regards it as Tertiary.

Thin sections were made from 92 rocks, and two specimens were chemically analyzed. Most of the laboratory work was done during the 8 months following the field work.

The writer wishes to express his indebtedness to Mr. W. S. Burbank for numerous helpful suggestions in the mapping of the area and in pointing out certain problems connected with it. The writer is furthermore indebted to Dr. H. L. Alling and Dr. Q. D. Singewald of the Department of Geology of the University of Rochester for assistance in the laboratory work and in consultation on various problems concerning this investigation.

TOPOGRAPHY

Stony Mountain, 12,698 feet high, is in an area of mountains many of which attain elevations exceeding 13,000 feet. Stony Mountain is characterized by bold cliffs and deep ravines commonly having vertical walls. Greatest relief within the mapped area is on the eastern side where the canyon of Yankee Boy Creek is nearly 2,000 feet below the summit of Stony Mountain.

GENERAL GEOLOGIC SETTING

In the immediate vicinity of Stony Mountain is an area of Tertiary volcanic rocks maintaining an average elevation of 11,800 feet. A few miles to the east, however, along slopes of the Uncompahgre Valley formations range in age from Algonkian to Recent. Rocks pertaining to the present study may conveniently be divided into three main groups: (1) Algonkian quartzite, (2) Miocene, bedded volcanic rocks several thousand feet thick, and (3) Upper Miocene or Pliocene plutonic rocks with associated igneous and clastic dikes. The last includes the Stony Mountain stock.

The Stony Mountain stock is part of a larger complex which also includes Mt. Sneffels, 2 miles northwest of the margin of Figure 1. A narrow connecting arm extends across the center of Yankee Boy Basin. Neither an adequate base map nor the time was available to study the Mt. Sneffels portion, but reconnaissance suggests that its igneous history probably was essentially the same as at Stony Mountain.

The portion of the stock south of the connecting arm is roughly circular, with greatest dimension about a mile in north-south direction. The vertical extent of the outcrops is 1900 feet. The surrounding volcanic rocks are remarkably undisturbed from their nearly horizontal attitude but locally they have been metamorphosed to a distance of several hundred feet. Apophyses and radial dikes extent from the main body; other dikes cut it. Even clastic dikes are associated. Perplexing, irregular-shaped outliers of many sizes occur in the bedded volcanic country rock, some too small to be mapped. Numerous younger veins cut the stock and the associated dikes.

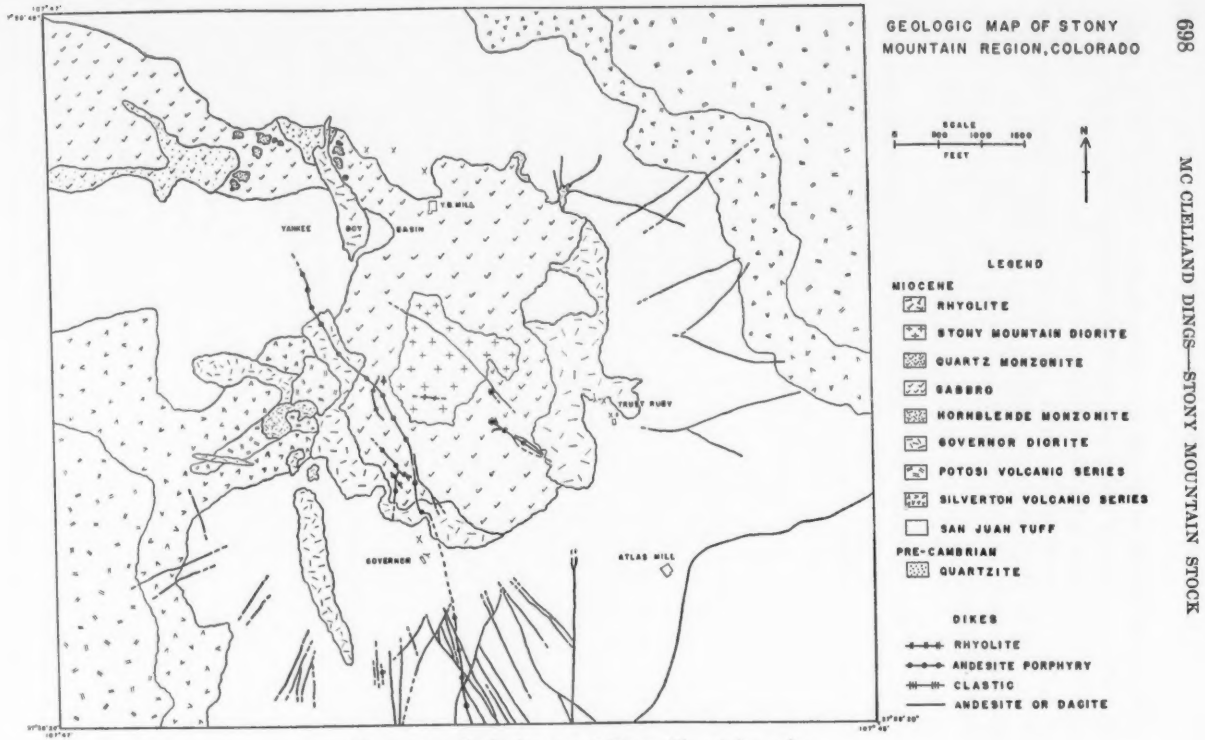


FIGURE 1.—*Geologic map of Stony Mountain region*

Showing mineral variations with changes of altitude as the border of the stock is approached.

In contrast to Cross' earlier generalization of a single gabbro-diorite mass with salic differentiation facies, the present detailed study indicates four closely related but separate instrusions, two of which have salic differentiation facies. The sequence from oldest to youngest was: "Governor" diorite, gabbro, "Stony Mountain" diorite, rhyolite. The first three types are separated not by gradational contacts, as believed by Cross, but by sharp contacts.

PRE-CAMBRIAN QUARTZITE

On the prominent ridge trending northwest from Yankee Boy Basin are numerous irregularly shaped patches of quartzite (Plate 1) that are distinctive lithologic features in this area of otherwise igneous and pyroclastic rocks. Cross (1899) mapped them as two large masses intercalated with rhyolite. His description and interpretation (p. 5), which differ markedly from the writer's, are:

> "The small body of upturned quartzites with an intercalated rhyolite sheet, occurring in Canyon Creek [Yankee Boy Creek], north of Stony Mountain, has been referred to the Algonkian, or oldest sedimentary group of rocks, because it is known, as stated in the introduction, that an immense series of quartzites, schists, slates, etc., does exist beneath the Paleozoic limestones near Ouray, projects up into the volcanic series at several places five miles or less to the eastward, and rises to elevations of over 14,000 feet in the Needle Mountains, on the southern border of the San Juan region. It is supposed that the quartzites of Canyon Creek represent a sharp pinnacle of the post-Cretaceous topography which was buried by the great volcanic accumulations of the San Juan.
>
> The quartzites of this area are rather coarse grained, grading into a fine conglomerate. They consist almost entirely of pale, smoky or somewhat clouded quartz grains, with only a small amount of kaolin in the interstices. The beds are massive banks, the bedding being, however, distinct, with variable strikes. The prevalent one seems to be about N.65°W., and the strata are in places nearly vertical.
>
> These quartzites, if not Algonkian in place, must be supposed to form from a block floated up in the gabbro magma of the Mt. Sneffels-Stony Mountain eruption. But since the mass projects on one side into the San Juan tuffs, and has but a narrow arm of the gabbro-diorite on the other side, it does not seem possible that a mass of this size can have been floated into this position."

The writer agrees with Cross in assigning the quartzite to the pre-Cambrian but differs by interpreting the rhyolite as part of the stock for reasons to be advanced. Apart from the age of the rhyolite, however, there is convincing evidence that the quartzite masses were carried up in the gabbroic magma. Detailed work revealed that, instead of forming two large masses, the quartzite actually exists in small isolated patches, some too small to be mapped, others up to 150 feet in length. Some clearly are separated by igneous rock, others by an unconsolidated mantle that conceals the bedrock relations. Many of the latter, nevertheless, though no more than 10 to 20 feet apart, differ decidedly in strike and dip. For example, immediately east of the rhyolite are two large blocks. One strikes due north and dips steeply west; the other

Showing mineral variations with changes of altitude as the border of the stock is approached.

strikes nearly due east and dips steeply south. Since there is no evidence of sharp folding or faulting within any one block, radically different dips cannot reasonably be ascribed to such causes. After careful examination of every quartzite outcrop, the writer found extreme variation in attitude rather than a prevailing N. 65°W. strike. The small size of even the largest blocks, the sharp divergence in attitude, and the presence of some undoubted quartzite inclusions favor the hypothesis that all are inclusions.

BEDDED VOLCANIC SERIES

GENERAL STATEMENT

The area contains three series of nearly horizontally bedded volcanic rocks: the San Juan tuff, Silverton volcanic series, and Potosi volcanic series. Their total thickness of approximately 2800 feet in the area mapped represents only a fraction of the total thickness in adjacent areas. They represent part of the volcanic material erupted in huge quantities during Tertiary time. Burbank (1932, p. 37) states that eruption reached maximum intensity during Miocene time but that it may have continued into the Pliocene. Portions of the series have been described in more or less detail in this and other parts of the San Juan Mountains; the reader is particularly referred to Cross and Larsen's (1935, p. 57-100) review of these complex volcanic rocks.

SAN JUAN TUFF (MIOCENE)

The lowest division within the area is the San Juan tuff which borders the stock almost everywhere except at the higher elevations on the west side of Stony Mountain. In places metamorphism extends hundreds of feet from the stock, and the tuff grades imperceptibly into the Governor diorite or the gabbro. In other places the tuff shows little metamorphism, and its contact with the stock is not difficult to establish.

The San Juan is a complex of tuffs, breccias, agglomerates, and conglomeratic beds, whose larger fragments are prevailingly fine-grained andesite and latite, the former being more abundant. The almost entire absence of rhyolite distinguishes it from the overlying Silverton series (Burbank, 1933, p. 140).

SILVERTON VOLCANIC SERIES (MIOCENE)

The Silverton volcanic series, unconformably overlies the San Juan tuff. It is essentially a complex of andesitic, latitic, and rhyolitic tuffs, breccias, and flows that reach a maximum thickness of about 3000 feet near Silverton (Cross and Larsen, 1935, p. 59). In the Stony Mountain region the maximum thickness is about 500 feet, and the most abundant rock type is a dark pyroxene andesite. Contact between the Silverton

FIGURE 1. STONY MOUNTAIN

FIGURE 2. QUARTZITE INCLUSIONS (LIGHT PATCHES IN LOWER CENTER) IN GABBRO

STONY MOUNTAIN AND QUARTZITE INCLUSIONS

FIGURE 1. TABULAR JOINTING IN RHYOLITE

FIGURE 2. CLIFFS OF STONY MOUNTAIN DIORITE

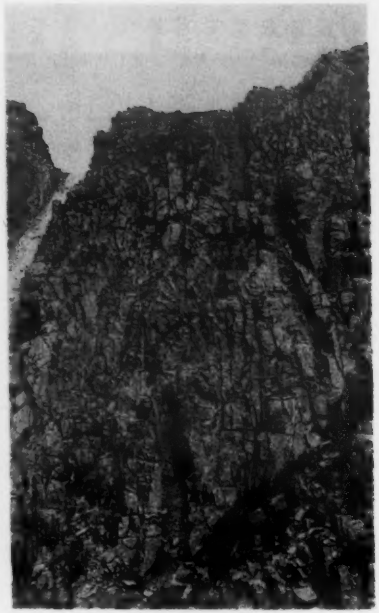


FIGURE 3. GABBRO
Rhyolite dike extends down from notch.

EXPOSURES OF GABBRO, DIORITE AND RHYOLITE

series and the stock can be seen only on the west side of Stony Mountain where the series has a maximum thickness of 310 feet.

POTOSI VOLCANIC SERIES (MIOCENE)

An erosional unconformity separates the Potosi volcanic series from the Silverton. Miocene age for the Potosi has been established by Emmons and Larsen (1923, p. 12) on the basis of plant remains found near Creede, Colorado. The Potosi is characterized by rhyolitic and latitic tuffs, breccias, and flows which contrast with the predominantly andesitic rocks of both the San Juan and the Silverton. The writer assigns all the Potosi rocks shown on Figure 1 to the "Treasure Mountain" quartz latite, one of six divisions established by Cross and Larsen (1935, p. 68).

Erosion has removed most of the Potosi rocks adjacent to the stock, but on Potosi Peak, only 6500 feet northeast of Stony Mountain, nearly 1300 feet of the series remains. The stock is in contact with the Potosi only on the western side of Stony Mountain (Fig. 1), where rhyolitic tuff and overlying quartz latite flow occupy the topmost 300 feet of two prominent ridges.

GOVERNOR DIORITE

GENERAL STATEMENT

The earliest rock constituting the stock is essentially a dark, fine-grained diorite, here termed the Governor diorite because it is typically exposed a short distance north of the Governor mine. The most prominent lithologic variation occurs along the west-central side of the stock where diorite grades into a light-colored, hornblende-bearing monzonite facies. This salic facies will be treated separately since it shows numerous interesting features pertaining to the problem of differentiation in igneous rocks. The diorite forms the outer border of most of the main stock and also occurs in disconnected, but closely associated, outlying bodies of various shapes and sizes (Fig. 1). The intrusion sent out numerous apophyses and radial dikes; some extend far up into the Potosi series, so intrusion was post-Treasure Mountain and is tentatively considered upper Miocene.

The vertical range of exposures is 1875 feet; the highest outcrop occurs about 275 feet northwest of the summit of Stony Mountain, and the lowest in the extreme southeast, near the junction of Governor and Yankee Boy creeks. In general the diorite is better exposed on the western than on the eastern side of the stock.

Most of the outer border of the diorite is in contact with the San Juan tuff. The intensity of contact metamorphism in the tuff ranges remarkably from place to place through a zone from a few feet to nearly 1000

feet in width. In many places, as the intrusion is approached, the metamorphic rock shows a perfect gradation from fragmental tuff into uniformly appearing igneous rock. The difficulty of attempting to delimit the two rock types in the field was soon observed, and it was necessary to form some criterion for separating them on the map. Near the intrusion the larger fragments of the tuff are conspicuous on weathered surfaces, even where the freshly broken surface resembles the usual fine-grained diorite. The contact was placed where the fragments disappear regardless of the fact that on the tuff side the rock often appears more igneous than fragmental.

In contrast to the usually wide metamorphic zones developed in the San Juan tuff, the Silverton and Potosi volcanic rocks generally show only slight metamorphism where intruded by the diorite. In a number of places the contact between diorite and Potosi quartz latite is sharp. The same applies to the Silverton pyroxene andesite, although the contacts are not so commonly exposed. With no apparent consistency, nevertheless, either the Silverton or Potosi rocks may be locally metamorphosed for several feet.

The varying width of the metamorphic zones shows that at different places the heat effects or the penetration by volatile constituents of the magma varied considerably. In general the metamorphic effects are much more pronounced in the more previous San Juan tuff than in the Silverton or Potosi rocks, and this suggests that permeability was the controlling factor.

Another important feature of the surrounding bedded volcanic rocks is their undisturbed, nearly horizontal attitude. In numerous places excellent vertical exposures along the contacts reveal practically no bending, fracturing, or warping of the pyroclastic rocks. This definitely suggests that emplacement was predominantly by some method other than by direct forcing apart of the intruded beds. The dip of the contact with the bedded volcanic rocks is nearly vertical in all places where it could be observed.

ROCK DESCRIPTION

Hand specimens of Governor diorite generally show uniform appearance regardless of location. The weathered surface ranges from dull gray to reddish brown and occasionally is pitted due to removal of large augite crystals. The fresh rock is fine-grained and greenish gray to greenish black. A few tablets of plagioclase and grains of augite, both averaging from 1 mm. to 2 mm. in length, may be seen imbedded in the nearly aphanitic groundmass. Exclusive of the salic differentiate, the most notable variation in the mass is a facies on the eastern and

northern sides of the stock showing much greater development of large augite crystals that occasionally reach 5 mm. in diameter but average about half this size.

Numerous thin sections show reasonably uniform texture throughout. The texture commonly is hypautomorphic with grains of widely varying sizes that commonly do not exceed 0.7 mm. in diameter. In general the larger crystals are plagioclase, which may or may not be zoned, and less abundant augite. Plagioclase, augite, and orthoclase are most abundant and are associated with varying amounts of hypersthene, quartz, pyrite, apatite, chlorite, calcite, biotite, and very small amounts of sphene, hematite, sericite, and epidote.

Plagioclase has formed in two generations. The first is labradorite, averaging $Ab_{43}An_{57}$; the second, andesine, averages $Ab_{58}An_{42}$, is considerably the more abundant, and occurs mainly as subhedral grains seldom longer than 0.4 mm. Albite twinning lamellae are generally sharp in labradorite but are poorly defined or absent in andesine. Zoned crystals are abundant, particularly in the first generation (Pl. 4).

Orthoclase occurs widely and fairly uniformly distributed in irregularly shaped grains of diverse sizes. It is usually quite fresh and relatively free from inclusions.

Augite is uniformly distributed in nonpleochroic, colorless to very pale-green, single grains or clusters. Individual grains range from minute specks to 1 mm. in diameter. Anhedral to euhedral crystals occur, with the former greatly predominating. The maximum extinction angle ($Z \wedge C$) is 43 degrees. A few large grains show diallage parting, and many others have twin seams developed with (100) as the twinning plane. Augite is often greatly altered to either chlorite and calcite or to a fine-grained aggregate of biotite and magnetite, the biotite showing further alteration to chlorite.

Light-green hypersthene is a minor constituent found only in a few thin sections. It occurs mainly as altered subhedral prisms slightly larger than those of augite, but there are rare euhedral crystals. The hypersthene is optically negative and has very faint, pink pleochroism which indicate a low iron content near enstatite in composition.

Original biotite, not a common constituent of the diorite, occurs in small extensively corroded grains that average from 0.1 to 0.2 mm. in diameter. It is strongly pleochroic with X = pale yellow, Y = light reddish brown, and Z = dark brown. Alteration commonly causes loss of color without loss of strong birefringence. Magnetite and apatite are the most common inclusions; commonly biotite encloses a single grain of irregularly rounded magnetite.

Quartz, in varying amounts, irregularly and intricately fills interstices. Many grains are easily overlooked except under very high magnification. Slight undulatory extinction and relative paucity of gas or fluid-filled cavities are characteristic. Microscopic intergrowths of quartz with orthoclase are rare.

Apatite is widespread in small, colorless to very light blue, needle-shaped crystals that commonly range from 0.01 to 0.04 mm. in length. It is included by all the other orthomagmatic minerals.

Magnetite occurs in grains ranging from specks to 0.35 mm. in diameter. Perplexing age relationships with the other minerals suggest three distinct periods of formation. The first generation forms irregularly rounded grains within augite and hypersthene. Many other grains, however, 0.1 to 0.35 mm. long, appear to have formed as an end phase or deuteric product. Gillson (1927, p. 23) gives two characteristics by which a deuteric mineral may be recognized; namely, "the grain does not fit into the crystal mesh of the rock, but lies in a haphazard or random manner without regard to the boundaries of adjacent minerals," and secondly, that "the grain is of irregular shape, and sends arms or finger-like branches out into other minerals; or it encloses or partly encloses several adjacent grains of different mineral species." Magnetite was not found definitely enclosing several adjacent mineral species, but all the other criteria were observed. Some additional tiny grains seem to be derived still later from augite.

Rare pyrite and hematite have random distribution, although few specimens failed to show some pyrite. Hematite borders the magnetite.

Every thin section contains considerable chlorite and calcite, and a few show minor development of epidote and sericite.

CHEMICAL AND MODAL ANALYSES

A chemical analysis (Table 1) of the Governor diorite was made from a specimen collected 290 feet due west of the summit of Stony Mountain at an altitude of 12,510 feet. Both megascopically and microscopically

TABLE 1.—*Chemical and modal analyses of the Governor diorite*

R. B. Ellestad, analyst, Rock Analysis Laboratory, University of Minnesota.

Chemical analysis

SiO_2	60.15	Na_2O	3.53
Al_2O_3	18.03	K_2O	2.76
Fe_2O_3	2.73	H_2O+	0.53
FeO	3.27	H_2O-	0.09
MgO	2.18	CO_2	0.19
CaO	5.78	TiO_2	0.65
		Total	99.89

Modal analysis (weight percentage)

Quartz	7.9	Magnetite	4.2
Orthoclase	15.1	Chlorite	7.4
Andesine (An_{42})	40.8	Other constituents—including biotite, hypersthene, apatite, calcite, hematite, sphene, sericite	5.3
Intratelluric labradorite (An_{57})	10.2		
Augite	9.1		
		Total	100.0

it appears to be typical of the rock mass and far enough removed from veins to be free from any late alteration.

A comparison of Table 1 with the average chemical analysis of diorite as given by Daly (1914, p. 26) shows that the Governor diorite has 3.28% more silica, 1.36% more alumina, and 1.99% less magnesia than average.

A calculation of the analysis to the C.I.P.W. standard normative minerals would give a total of about 14% quartz, whereas in reality less than 8% is present.

Table 1 also tabulates a modal analysis (weight percentage) of the same specimen that was chemically analyzed. It can be noted from the mineral composition that the term diorite is merely one of convenience, for the rock contains too much orthoclase and quartz to be a normal diorite.

HORNBLENDE MONZONITE FACIES

General statement.—The most outstanding differentiation facies of the Governor diorite is a cupola of hornblende monzonite exposed at an average altitude of 12,400 feet along the prominent south-west-trending ridge near the summit of Stony Mountain. The greatest length of the monzonite body is about 700 feet; the width varies from 100 to 300 feet. An embayment coinciding with a small topographic saddle roughly divides the body into two equal parts (Fig. 1). Contacts with the Silverton and Potosi series, though not abundantly exposed, are sharp and nearly vertical.

Typical transition toward diorite is illustrated by a series of specimens collected across the hornblende monzonite body, southwest of its central embayment, along the crest of a ridge whose elevation does not vary more than 40 feet. Changes toward the border are essentially the same in each direction from the center. Most conspicuous are a decrease in hornblende and corresponding increase in augite. In addition, intratelluric labradorite becomes slightly more abundant, more glomeroporphyritic, and less corroded; late generation feldspar more calcic; and epidote less abundant as the border is approached.

Rock description.—Prominent megascopic features of the monzonite are the light-greenish-gray matrix and the conspicuous hornblende blades ranging from 2 to 4 mm. in length. Chlorite, epidote, and altered grains of feldspar can generally be readily distinguished.

The most notable microscopic features of the monzonite, as contrasted with the diorite, are the abundance of narrow hornblende blades (Pl. 4), the more sodic composition of the plagioclases, the small amount of augite, the greater development of epidote and sericite, and the total absence of hypersthene.

Hornblende occurs in unoriented, corroded blades some of which are twins having (100) as the twinning plane. The maximum extinction angle ($Z \wedge C$) is 17 degrees; cleavage is well developed; and pleochroism is strong, with X = yellowish green, Y = olive green, and Z = dark green. The hornblende apparently crystallized directly from the magma, for neither augite, partly replaced by hornblende, nor pseudomorphs of hornblende after augite were found.

Plagioclase of the monzonite occurs, as in diorite, in two distinct generations. The earlier is labradorite ($Ab_{43}An_{57}$) mostly forming anhedral to subhedral grains 1 to 2 mm. long. Albite twinning is well developed and in a few grains is combined with Carlsbad. Borders commonly are partly resorbed, and entire grains are injected by later oligoclase. The later generation, a sodic oligoclase ($Ab_{86}An_{14}$), is the most abundant feldspar. Individual crystals, seldom exceeding 0.8 mm. in length, are subhedral to anhedral.

The optical properties of the other minerals present (Table 2) are identical with those in the diorite.

Chemical and modal analyses.—The specimen selected for chemical and modal analyses (Table 2) represents the most salic type rich in hornblende.

A comparison of the chemical and modal analyses of the diorite and monzonite (Tables 1, 2) indicates both similarities and differences.

Total silica is approximately the same. The monzonite has nearly 2.5% less Al_2O_3 than the diorite; this can be explained chiefly by the greater proportion of Al_2O_3 in the more calcic feldspar of the diorite, in spite of differences in the total quantity of plagioclase. Slight

TABLE 2.—*Chemical and modal analyses of the hornblende monzonite facies of the Governor diorite*

R. B. Ellestad, analyst, Rock Analysis Laboratory, University of Minnesota.

Chemical analysis

SiO_2	60.23	CO_2	0.44
Al_2O_3	15.56	TiO_2	0.79
Fe_2O_3	3.20	ZrO_2	0.00
FeO	3.08	P_2O_5	0.34
MgO	2.73	S	0.01
CaO	4.21	Cr_2O_3	0.00
Na_2O	4.43	MnO	0.14
K_2O	3.42	BaO	0.11
H_2O+	1.10	Cl	0.03
H_2O-	0.10	F	0.00
		Total	99.92

Modal analysis (weight percentage)

Quartz	8.1	Augite	4.3
Orthoclase	18.7	Magnetite	5.8
Oligoclase (An_{14})	31.3	Other constituents—including chlorite, epidote, calcite, pyrite, sphene, apatite, hematite, sericite	13.1
Intratelluric labradorite (An_{57})	8.1		
Hornblende	10.6		
		Total	100.0

differences between the FeO and Fe_2O_3 content of the two rocks can best be attributed to the varying and uncertain amounts of iron contained in chlorite, augite, and hornblende, even though the modal amounts of magnetite do differ about 1.5%. The increase in Na_2O in the monzonite is due to more sodic feldspar which in turn is less abundant than the andesine of the diorite; hence the difference is not so great as the compositions of plagioclases would at first suggest. The difference in the K_2O content is not great and is accounted for chiefly by the relative differences in the amounts of orthoclase, since biotite is not sufficiently abundant in the diorite to affect the analysis materially. The greater

amount of water expelled above 110°C. in the monzonite is assumed to have come chiefly from hornblende. The CO_2 content is greater in the monzonite than the diorite, due to greater abundance of calcite.

Theory of origin.—Little or no gravitative settling could have taken place within the Governor diorite after emplacement, for it is reasonably uniform in composition and texture throughout both lateral and vertical extent. Fluctuations in equilibrium during crystallization, however, are evidenced by uniformly distributed oscillatory zoning in plagioclases. Intratelluric crystals of plagioclase show no variation in distribution in a vertical range of 1800 feet.

The complete gradation from Governor diorite to hornblende monzonite, in contrast to the sharp contacts with other rocks, strongly suggests that the monzonite is a salic differentiate formed in situ. The evidence as to absence of crystal settling elsewhere, the existence of approximately the same quantity of intratelluric plagioclase in the monzonite as in the diorite, and the lower elevation of the monzonite than much of the diorite seem to preclude explaining the differentiation as due to gravitative settling after emplacement.

Transfer of material after emplacement but prior to solidification is the most reasonable theory to account for the hornblende monzonite. The transfer was probably accomplished by volatile constituents, since the hornblende monzonite occupies a cupola where gases might be expected to collect. This explanation accounts for the many similarities and also for the differences, notably (1) greater Na_2O, K_2O, and H_2O content and lesser Al_2O_3 and CaO content, due to differences in composition of plagioclases and in proportions of hornblende to augite, and (2) greater magmatic corrosion of early plagioclase in the hornblende monzonite than in the diorite.

In view of the evidence for emplacement by magmatic stoping provided by the undisturbed attitude of the volcanic rocks and the fact that the parent magma of the monzonite facies invades the relatively salic Potosi quartz latite, the change in composition might be ascribed to material incorporated from the wall rock. At no other place where diorite invades the same rock, however, is the monzonite facies developed. Moreover, neither included fragments of quartz latite nor segregations that could be interpreted as former inclusions could be found within the hornblende monzonite.

GABBRO

GENERAL STATEMENT

By far the greatest volume of rock composing the stock is a medium to dark gabbro locally showing great variation in composition, texture,

and color. The most prominent variation is a quartz monzonite facies mapped separately (Fig. 1) in the connecting arm northwest of Yankee Boy Basin. In addition, light-colored diorite grading into quartz diorite forms the bulk of the floor of Yankee Boy Basin and the cliffs east of Yankee Boy mill.

The gabbro is younger than the Governor diorite, for stringers extend into the diorite, and inclusions of diorite occur near the contact. In general the contact is sharp.

Contact of gabbro with San Juan tuff may be sharp or gradational, but, where sharp, the tuff had previously been metamorphosed by the earlier Governor diorite.

The metamorphic zone between gabbro and tuff varies from 20 feet on the northern and western sides to nearly 1000 feet on the northwestern border. The wider zones may represent places where the magma chamber spread out beneath the tuff, but no field evidence could be found to support this supposition. Differences in composition and extent of jointing in the intruded rock might result in greater local permeability causing marked differences in the intensity of metamorphism. However, this does not seem to account solely for the great variations that exist, and an additional factor has probably been the irregular concentration of volatile and less viscous parts of the magma.

ROCK DESCRIPTION

Although the bulk of the mass is essentially a gabbro, no single specimen is typical of the entire stock. The rock varies from noritic gabbro to quartz monzonite yet it possesses certain characteristic petrographic features—dark-green color, medium grain size, and diabasic texture.

Microscopic examination shows texture variations from diabasic to hypautomorphic. The most abundant minerals are plagioclase and augite, with associated orthoclase, hypersthene, biotite, quartz, apatite, chlorite, calcite, white mica, hematite, pyrite, leucoxene, and sphene.

Plagioclase, averaging $Ab_{47}An_{53}$, commonly occurs as subhedrons. Albite twinning is prominent; Carlsbad combined with albite twinning is rare. Zoning becomes increasingly abundant toward the border of the mass.

The orthoclase content ranges from 4 to 27 per cent by weight. It is characteristically late as evidenced by irregularly shaped grains including subhedral or euhedral grains of plagioclase, augite, and most of the accessory minerals.

Pale-green augite occurs in subhedral and, less commonly, anhedral grains of many different sizes, some as long as 5 mm. Many crystals show seamlike twinning with (100) as the twinning plane; rarely herringbone structure is developed where (100) twins combine with polysynthetic (001) twins. The extinction angle ($Z_A C$) is 43 degrees, and occasionally a diallage parting is developed. Augite is frequently intergrown with plagioclase, and either one may include the other.

Hypersthene varies greatly in amount and is completely absent in some facies. The size and shape of grains differ, but many are subhedrons, which commonly are corroded. The hypersthene is very faintly pleochroic and biaxial negative; hence, it probably is low in iron and near enstatite. Small euhedral crystals of plagioclase are common inclusions.

Biotite occurs in almost all specimens in different-sized grains that quite consistently show considerably corroded edges and likewise partial or almost complete alteration to chlorite. The mica is strongly pleochroic, with X = light yellow, Y = faint reddish brown, and Z = deep brown. A very small amount of secondary biotite is associated with and derived from augite.

TABLE 3.—*Modal analyses of a series of gabbro specimens*

Showing mineral variations with changes of altitude as the border of the stock is approached.

Specimen no.	Altitude	Plagioclase	Orthoclase	Quartz	Total of: augite, hypersthene, biotite, chlorite, calcite, sericite	Magnetite	Total minor accessories: apatite, sphene, hematite, pyrite, leucoxene
36D69	12698	47.1 (An_{51})	12.8	1.3	32.5	5.6	0.7
36D319	12545	49.5 (An_{49})	5.2	0.5	34.8	8.2	1.8
36D314	12450	47.9 (An_{50})	9.9	1.6	29.4	9.5	1.7
36D312	12350	48.2 (An_{62})	6.7	1.4	30.5	11.3	1.9
36D309	12155	53.4 (An_{57})	8.3	0.2	28.8	8.5	0.8
36D147	11850	50.8 (An_{49})	13.3	8.1	22.8	3.2	1.8
36D143	11465	55.3 (An_{55})	12.3	0.6	25.7	5.3	0.8
36D139	11140	45.5 (An_{47})	14.9	8.7	21.9	7.1	1.9
36D105	10870	44.6 (An_{46})	15.1	8.8	21.6	8.2	1.7

Some quartz was found in every rock, although only a few isolated grains are present in some. It partially penetrates and fills interstices between the other minerals. Small gas- or liquid-filled cavities are rare, and only a moderate amount of undulatory extinction can be observed.

There seem to be three generations of magnetite. Early magnetite occurs in anhedral to subhedral crystals included by other minerals. Other grains of magnetite have formed during alteration of the mafic minerals and are found bordering the mafic minerals or else as irregular small grains confined to the host mineral and associated with much calcite, chlorite, or bleached mica. Most of the magnetite, however, occurs as irregularly shaped blotches that penetrate all the other minerals.

Apatite, a consistent and widespread accessory mineral, occurs most commonly in euhedral needles averaging less than 0.3 mm. long.

Sphene, rarest of the accessory minerals, occurs in small, irregular, wedge-shaped grains.

MODAL ANALYSES

Modal analyses (Table 3) were made of nine specimens collected along a southeasterly line extending from the top to the base of Stony Mountain. The series of specimens represents a vertical range of 1828 feet and an approximate horizontal distance of 2500 feet.

The analyses bring out the mineral variations of the mass. It can be noted that the composition of plagioclase varies with no general consistency as the bottom of the stock is approached except that near the border the plagioclase becomes more sodic and is associated with more orthoclase and quartz. This is not a perfect gradational relationship as a comparison of specimens 36D147 and 36D143 will indicate. The

plagioclase of specimen 36D147 has a composition of andesine and is associated with a moderately high amount of quartz and orthoclase, but specimen 36D143, which is lower and also nearer the border of the mass, shows plagioclase of labradorite composition associated with less than 1 per cent quartz.

TABLE 4.—*Chemical analysis of the gabbro from the summit of Stony Mountain*

L. G. Eakins, analyst. From the Telluride folio (No. 57) p. 7

SiO_2	52.05	MgO	5.03
Al_2O_3	17.96	K_2O	1.61
Fe_2O_3	4.09	Na_2O	2.99
FeO	6.33	H_2O	0.97
MnO	0.43	P_2O_5	0.31
CaO	8.64		
		Total	100.41

The apparent variation in magnetite content is fortuitous, for in numerous places the rock grades horizontally within a few hundred feet from one rich in magnetite to one very low in magnetite, although not even the most salic phases show a complete absence of magnetite.

CHEMICAL ANALYSIS

A chemical analysis (Table 4) of gabbro from the summit of Stony Mountain was published by Cross (1899) in the Telluride folio. The analysis is listed by Washington (1917, p. 533) as a superior analysis of a fresh rock with a rating of A2.II.

Table 5 compares the calculated norm of the chemical analysis with the author's modal analysis of a specimen from this general area. The greatest difference between the two rocks lies in the percentage of orthoclase, as the writer's specimen was computed to contain over 3 per cent more orthoclase than the one analyzed. The apparent discrepancy between the two plagioclases is readily accounted for by the presence of considerable augite in the mode, which would reduce the CaO of the norm and subsequently reduce the relative abundance of anorthite. The difference between the two quantities of hypersthene can likewise be explained by assuming that part of the MgO of the mode is present in chlorite and biotite which are not standard normative minerals. Any discrepancies between the two analyses can be accounted for by the uncertainty of composition of such minerals as augite, chlorite, and biotite, or else the original composition variation of the two specimens analyzed could in itself readily account for the differences.

QUARTZ MONZONITE FACIES

The quartz monzonite facies crops out over two fairly large areas in the connecting arm from Yankee Boy Basin towards Mt. Sneffels (Fig 1). The gabbro commonly grades imperceptibly into quartz monzonite, but

TABLE 5.—*Norm and mode of two different specimens of gabbro from the summit of Stony Mountain*

C. I. P. W. standard normative minerals*		Modal analysis of specimen 36D69**	
			vol. %
Anorthite	33.36	Labradorite (An_{51})	47.1
Albite	25.15	Augite	15.1
Hypersthene	18.72	Orthoclase	12.8
Orthoclase	9.45	Hypersthene	12.2
Diopside	6.12	Magnetite	5.6
Magnetite	5.80	Chlorite, calcite, sericite	3.1
Quartz	0.96	Biotite	2.1
Apatite	0.67	Quartz	1.3
		Minor accessories: mostly apatite	0.7
Total	100.23	Total	100.0

* Norm accompanying the chemical analysis given as Table 4.
** No chemical analysis made of this specimen.

in some places irregularly shaped apophyses of the quartz monzonite definitely penetrate the gabbro with parts of the apophyses showing sharp dikelike contacts.

The quartz monzonite is medium- to fine-grained and differs very little in average grain size from the adjoining gabbro. Its color ranges from light pinkish gray to light greenish gray but becomes a much darker green as it grades into the gabbro.

ORIGIN OF QUARTZ MONZONITE

The dikes and apophyses indicate that the quartz monzonite is slightly later in time of emplacement than the gabbro. The presence of both gradational and sharp contacts suggests that the quartz monzonite magma was forced upward into the hot, but partially solidified, gabbro. Differentiation must have occurred at depth, therefore, and not in situ as postulated for the hornblende monzonite derived from the Governor diorite.

Field and microscopic data yield only negative evidence as to the mechanism by which the quartz monzonite originated. As the gabbro

penetrated well into the Potosi rhyolites and latites, salic rocks must have been added to the magma, yet none partly assimilated can be found. Moreover, no direct evidence can be set forth to support a theory of fluid or gaseous transfer. It is very possible that fractional crystalliza-

TABLE 6.—*Average mode (weight percentage) of three specimens of the quartz monzonite facies of the gabbro*

Quartz	17.6	Sericite	5.6
Orthoclase	22.1	Calcite	2.4
Andesine (An_{36})	35.8	Epidote	2.1
Magnetite	2.9	Minor constituents—including apatite, pyrite, biotite, hornblende, sphene, hematite, leucoxene	4.2
Chlorite	7.3	Total	100.0

tion at depth by crystal settling or "filter press" mechanism may be the answer.

STONY MOUNTAIN DIORITE

GENERAL STATEMENT

The Stony Mountain diorite, vertically exposed for 1320 feet, is entirely restricted to the east side of Stony Mountain and is surrounded by gabbro. Jointing is prominent (Pl. 2), tends to be very irregular, and is more closely spaced than in the gabbro. Chemical weathering has progressed along the joints to a moderate extent with notable development of hydroxides of iron.

The Stony Mountain diorite contains inclusions of gabbro and sends apophyses into the gabbro. Furthermore, small granite aplite dikes found occasionally in the gabbro end abruptly against the diorite. These evidences of age relationship between the two rocks are scanty, for a sharp contact can usually be followed for great distances without obtaining any definite clue as to which is younger. The relations cited, however, clearly prove that the Stony Mountain diorite is younger than the gabbro, which in turn is younger than the Governor diorite.

ROCK DESCRIPTION

A noteworthy feature of the diorite is its uniform megascopic appearance. The typical rock is fine-grained, dark greenish black, and breaks with a subconchoidal fracture. A faint diabasic texture may often be ob-

served on the fresh surface, and still more frequently on the slightly weathered surface.

There are no large differentiation facies such as occur in the Governor diorite or in the gabbro, and the minor facies are small, gradational patches only slightly different from the normal type. No consistent border features have developed, but in a very few places an increase in biotite over augite can be noted. Some of the minor facies may be medium-grained or almost aphanitic, the latter being the more common.

In thin section the typical diorite is seen to be uniformly fine-grained with a texture ranging from hypautomorphic to diabasic, the former much more common.

The principal plagioclase is andesine ($Ab_{59}An_{41}$), occurring commonly as subhedral laths averaging 0.3 mm. in length. Albite twinning is very prominent; a few crystals show pericline twinning, others a combination of albite and Carlsbad twinning. The few inclusions present are small anhedral grains of magnetite and augite as well as euhedral needles of apatite.

In addition to the main plagioclase there are a few moderately corroded intratelluric crystals of a more calcic andesine ($Ab_{54}An_{46}$) averaging 0.6 mm. in length.

Orthoclase averaging 0.3 mm. in length fills the interstices between andesine and augite or may even enclose them.

Augite ranges from minute specks to 0.6 mm. in diameter; the average is 0.2 mm., and grains smaller than this are more common than larger ones. Anhedrons greatly predominate, in contrast to the more perfectly crystallized augite of the gabbro. The maximum extinction angle measured from the slow ray is 43 degrees. Anhedral magnetite and euhedral apatite, the former predominant, are abundant as inclusions.

Brown biotite, found in a few border facies, occurs as irregularly shaped grains averaging 0.4 mm. in diameter. Biotite includes small crystals of augite, magnetite, andesine, and apatite.

Magnetite occurs in small anhedral grains averaging 0.3 mm. in diameter. A few show slight border alteration to hematite. Some slightly larger grains appear to be of end phase origin, and some magnetite has also resulted locally from alteration of augite and biotite.

MODAL ANALYSES

Modal analyses (Table 7) of three specimens from widely separated localities and different elevations depict the most extreme mineralogic differences that may be encountered. Specimen 36D163 represents the typical diorite. Specimen 36D310 was collected near the border and represents a minor biotite-rich facies. Specimen 36D57 is one of the darker, almost aphanitic facies. Table 7 shows that the rock does not vary to any unusual degree even in its most extreme types. The plagioclase content varies less than 4 per cent. Orthoclase is nearly constant; it shows about the same abundance as in the gabbro. Augite varies within 11 per cent; it can be noted that a decrease in augite is compensated by a proportional increase in biotite, and the total of the two is very similiar in all three specimens. Magnetite, present in moderate amounts, is not so abundant as in the gabbro. Hypersthene is not present.

The Stony Mountain diorite shows a marked similarity to both the Governor diorite and the gabbro in regard to essential minerals, the late occurrence of orthoclase, and the consistent presence of magnetite and

apatite. It is reasonably assumed that the Stony Mountain diorite magma was closely related to the magmas of the Governor diorite and gabbro even though a period of time intervened to permit the solidification of each earlier intrusion.

TABLE 7.—*Modal analyses (weight percentage) of three specimens of the Stony Mountain diorite*

	Specimen 36D310	Specimen 36D163	Specimen 36D57
Quartz	5.9	4.4	3.3
Orthoclase	13.2	15.2	14.8
Plagioclase	45.1 (An_{40})	41.4 (An_{42})	42.4 (An_{41})
Intratelluric plagioclase	4.3 (An_{48})	4.1 (An_{49})	5.1 (An_{48})
Biotite	15.1	3.2	0.8
Augite	7.5	14.8	18.5
Magnetite	3.4	5.7	5.1
Chlorite	2.1	7.1	6.1
Minor constituents—including apatite, calcite, sericite, hematite, pyrite	3.4	4.1	3.9
Totals	100.0	100.0	100.0

RHYOLITE

GENERAL STATEMENT

The last large intrusion in the stock was a rhyolite forming a mass about 1500 feet long and 300 feet wide. Excellent exposures along Yankee Boy Creek show prominent vertical flow structure developed at the eastern end, and a short distance farther west the rock is tabularly jointed (Pl. 2). The rhyolite contact with gabbro can be seen at an elevation of 11,525 feet along a tributary stream.

ROCK DESCRIPTION

The rhyolite is light gray to pale pink on a fresh surface, becomes soft and chalky white adjacent to the numerous veins, and weathers dull gray or light reddish brown. Quartz and orthoclase phenocrysts, mostly euhedral, each average 2 mm. in diameter. Quartz tends to stand out on a weathered surface, where crystal faces may be well preserved.

As seen in thin section, the rhyolite is very felsic, with the mafic minerals constituting only about 5 per cent by weight. Phenocrysts of orthoclase and quartz (Pl. 4), the former slightly more abundant, constitute about 16 per cent. The feldspar is usually more or less banded by small dustlike inclusions.

The groundmass is principally orthoclase and quartz in anhedral grains ranging from 0.02 to 0.05 mm. in diameter. Hornblende and biotite occur very sparingly in

FIGURE 1. CONTACT OF GABBRO (LEFT) AND STONY MOUNTAIN DIORITE (RIGHT)

FIGURE 2. CONTACT OF RHYOLITE AND HIGHLY METAMORPHOSED TUFF

CONTACTS

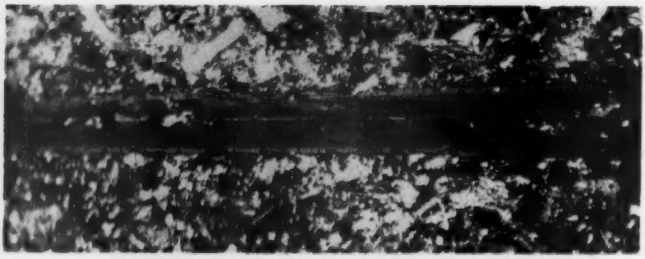


FIGURE 1. HORNBLENDE BLADE—HORNBLENDE MONZONITE FACIES OF DIORITE. X 32.

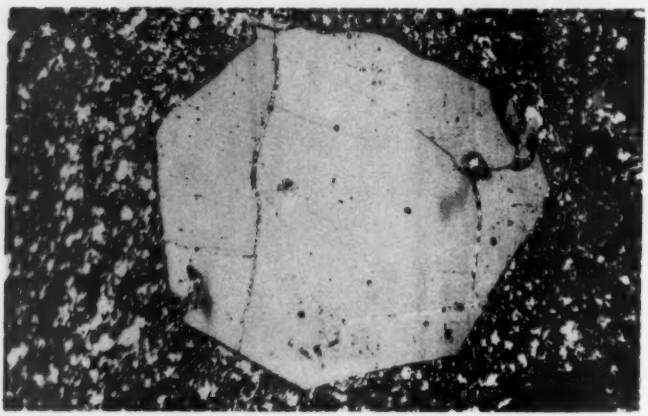


FIGURE 2. PARTIALLY CORRODED QUARTZ PHENOCRYST OF RHYOLITE. X 32.

FIGURE 3. ZONED PLAGIOCLASE CRYSTAL OF GOVERNOR DIORITE. X 90.

PHOTOMICROGRAPHS

subhedrons less than 0.05 mm. in size. Small grains and needles of apatite are found only as inclusions in orthoclase and quartz. Tiny grains of rounded zircon and of subhedral sphene are sparse. Magnetite forms irregular grains ranging from 0.01 to 0.03 mm. in diameter.

AGE OF RHYOLITE

Cross (1899, p. 2) assigned the rhyolite an Algonkian age because (Algonkian) quartzites are intercalated with the rhyolite in Canyon Creek (Yankee Boy Creek on later maps) and because the rhyolite does not penetrate the San Juan tuff. Evidence contrary to both these statements was found during the detailed study made by the present writer.

Although the quartzite masses do occur near the rhyolite, there is absolutely no evidence that the two are intercalated. Actually, only one quartzite block is in contact with the rhyolite, this being on the northwest side. Most of the remaining quartzite blocks are entirely surrounded by gabbro, and some lie as much as 1000 feet from the rhyolite.

The northernmost exposure of rhyolite definitely penetrates the San Juan tuff. This conclusively proves the rhyolite to be at least Tertiary. In addition there is indirect evidence that the rhyolite is younger than either the gabbro or the Stony Mountain diorite. One bit of evidence is found at the eastern contact of rhyolite and metamorphosed tuff along Yankee Boy Creek (Pl. 3) where gabbro has intensely metamorphosed the San Juan tuff but has not affected the adjoining rhyolite, so the rhyolite must have been intruded subsequent to the gabbro. Other evidence is based upon rhyolite and associated pitchstone dikes which extend southward from the main mass. The dikes include gabbro, Stony Mountain diorite (Fig. 1) and also San Juan tuff at a distance of nearly 2000 feet beyond the southernmost border of the stock.

DIKES

GENERAL STATEMENT

Many igneous dikes of various sizes and compositions are genetically related to the Stony Mountain intrusion. By far the largest number are andesites that radiate from the main intrusion (Fig. 1). In addition to the igneous dikes, two clastic dikes penetrate the gabbro and Governor diorite. The dikes can be divided into five main divisions, which in probable order from oldest to youngest, are: "radial" andesite dikes; aplites associated with the three basic intrusions; clastic dikes; "non-radial" andesite, dacite, and andesite porphyry dikes; and rhyolite and pitchstone dikes.

"RADIAL" ANDESITE DIKES

The radial andesite dikes border the stock chiefly on its northeastern, eastern, and southern sides. Those to the northeast clearly radiate from the Governor diorite, and the position of the others is very significant. A

study of the geologic map (Fig. 1) shows that practically all, if extended toward the stock, would come from this early diorite, although rock mantle now obscures their inward extension. Of equal or greater importance is the fact that no dikes of this character penetrate the gabbro or the Stony Mountain diorite. Hence, the dikes must be associated with the Governor diorite period of intrusion.

Atlas Flat, which lies south of the stock, contains a maze of dikes whose positions can be fairly well traced by numerous outcrops. They range in width from 2 to 60 feet, but most are about 10 feet wide. Vertical to nearly vertical dips prevail; an outstanding exception occurs near the Atlas mill, where a large dike dips 30° NW.

East of the stock, a number of large dikes are exposed along the steep slopes of Potosi Peak. Many are closely associated with and intensely altered by later veins. Since the altered material is easily eroded, most dikes now occupy ravines. The eastern dikes are exposed over a maximum vertical range of 1000 feet. Their width and composition are about the same as those in Atlas Flat. Most of them pinch out within a distance of 2500 feet.

A branching group of dikes or apophyses is well exposed in the sharp ravines and cliffs about 1300 feet due north of Yankee Boy Falls. One of the bodies shows all transitions from normal Governor diorite in the connecting trunk (apophysis) outward into an andesite dike having the same composition and texture as all the other radial dikes. Small stringers of bordering gabbro penetrate the lower part of the diorite trunk, so the gabbro is later than the dike.

Dikes on the steep slopes of the large basin west of the Governor mine are well exposed. All are andesites that become thinner away from the stock and so are assumed to be radial and genetically related to the Governor diorite.

Megascopically, the rocks composing the dikes show great similarity. The vast majority are aphanitic, without conspicuous phenocrysts; they are typically medium to dark gray, though some have a green tone. There is little vertical or horizontal variation within a dike. A few specimens from widely separated localities were examined in thin section. An intermediate andesine occurs in subhedral to euhedral forms. Augite, in subhedral forms, is the most abundant mafic mineral, although locally hornblende may be present in considerable quantity. Minor amounts of quartz and orthoclase are local, and euhedral needles of apatite have the same abundance as in the stock. Magnetite of late origin is always present. Hydrothermal alteration and weathering have developed calcite and chlorite in large quantities; the mafic minerals have largely been altered beyond definite recognition.

APLITE DIKES

A few aplite dikes occur in the Governor diorite, gabbro, and Stony Mountain diorite, but they are absent in the rhyolite. They are not common, however, to warrant detailed study. Nearly all are less than 3 inches wide and extend for only short distances. The vast majority are granite aplites having a light- to medium-pink color and a sugary texture, but a few light- to dark-gray, sugar-texture dikes were found that probably range in composition from monzonite to quartz diorite. It is notable that no pegmatite dikes were found in any of the rocks composing the stock.

CLASTIC DIKES

Two clastic dikes are intermittently exposed on the eastern side of Stony Mountain (Fig. 1). The width of the southern one averages 8 feet but attains a maximum of 16 feet, whereas the maximum width of the northern one is only 2 feet. The southern dike also contrasts with the northern in being much better exposed and locally associated with a dacite dike. The dikes dip steeply but have numerous local variations both in strike and in dip. From each main dike extend numerous branches, some of which rejoin. Near the western extremity of the southern dike an excellent vertical exposure shows a large branch of the dike pinching out in an upward direction.

The fragments composing the clastic dikes have a wide range in size, shape, and composition. The size ranges from microscopic to 14 inches in diameter, but fragments one inch or less in diameter predominate. The prevailing shapes are subangular to fairly well rounded, although some are either angular or almost spherical. Igneous material including andesite, diorite, dacite, monzonite, and granite predominates. There are also a few fragments of intensely weathered slate and of quartzite; the latter closely resembles the pre-Cambrian quartzite exposed near Ouray.

The typical rock is dull greenish brown or brown. Light-colored fragments stand out prominently in the more intensely altered, fine-grained matrix. Even under the microscope, the bonding material appears only as a clouded film.

The lithologic character indicates that material in the dikes was derived from older and lower-lying conglomeratic beds. The upward pinching of a branch of one of the dikes eliminates the possibility of fractures being filled from above. No quartzite beds are known to exist in this general locality either higher stratigraphically or in altitude than the stock. The most logical source would be the basal conglomeratic beds of the San Juan tuff and the underlying Telluride conglomerate which are composed of rocks similar to those found in the clastic dikes (Cross, 1935, p. 47, 57).

The writer believes that material now forming the clastic dikes was brought up by local explosive volcanic action having a force great enough to bring up the poorly consolidated material of the underlying beds. It is notable that the southern dike is closely associated with a dacite which in places shows the clastic material in an igneous matrix. The partial association of the igneous and clastic dikes may be due to a later origin of the dacite dike but it also suggests that the explosive action may have occurred just prior to the upward movement of the dacitic magma.

"NON-RADIAL" ANDESITE PORPHYRY, ANDESITE, AND DACITE DIKES

A group of seemingly related dikes ranges in composition from andesite to dacite and in texture from even-grained to porphyritic; they are not numerous, and only a few may be followed for any considerable distance. Although the average width is about 3 feet, there actually is great variation in both width and attitude. All the dikes send off minor branches, some of which rejoin.

The dikes of this group are younger than the "radial" andesite dikes. Not only do many of them transect Governor diorite, gabbro, and Stony Mountain diorite, but on the north side of Atlas Flat an andesite porphyry cuts one of the radial dikes. On the other hand, an andesite porphyry on the south slope of Stony Mountain is cut by a rhyolite dike that is believed to be essentially contemporaneous with the rhyolitic facies of the stock. This andesite-dacite group, therefore, is younger than the Stony Mountain diorite and probably older than the rhyolite.

RHYOLITE DIKES

Extending in a general southward direction from the main rhyolite intrusion of Yankee Boy Basin is a series of disconnected rhyolite dikes that penetrate all the rocks of Stony Mountain and even extend 2000 feet south of the stock into the San Juan tuff (Fig. 1). Most are tubular-shaped, but some occur as small isolated bodies that are cigar-shaped in ground plan. The latter vary from 1 foot to 12 feet in width at their centers; they may be composed of rhyolite, very resinous brown pitchstone, or a combination of rhyolite and pitchstone. Where together, either rhyolite or pitchstone may occupy the center.

In spite of minor variations in strike, the prevailing trend of the dikes is nearly north-south, parallel to the general belt in which they occur. The belt if extended northward would meet the main rhyolite body of the stock. Furthermore, the predominating dike rock has no phenocrysts but otherwise is similar to the rhyolite of Yankee Boy Basin. These features strongly suggest a genetic relation between the dikes and

the rhyolite facies of the stock, and field relations prove the dikes to cut late andesite porphyry as well as all the basic plutonic facies.

SUMMARY

The Stony Mountain stock has penetrated Miocene volcanic rocks belonging to the San Juan, Silverton, and Potosi series. Field and microscopic evidence reveal four distinct intrusions. The rocks are essentially diorite, gabbro, and rhyolite, although many local gradational facies were developed. The first intrusion is represented by a fine-grained, dark diorite (the Governor diorite) and numerous radial andesite dikes. An associated hornblende monzonite facies is believed to have formed by gaseous or liquid transfer of material into the previously emplaced dioritic magma.

The second intrusion is essentially a medium-grained, dark-green gabbro that varies considerably in composition within short distances. Two moderately large areas of quartz monzonite associated with the gabbro differentiated at depth prior to emplacement into the hot, partially solidified gabbroic mass. Blocks of pre-Cambrian quartzite, ranging from small fragments to 150 feet in length, were carried up by the gabbroic magma.

Clastic dikes containing fragments of lower-lying conglomeratic beds transect the gabbro and Governor diorite and suggest that very minor explosive activity occurred after the gabbro was formed; available evidence does not necessarily restrict the time of activity as prior to the next intrusion.

A third and much smaller intrusion followed the gabbro. The rock (Stony Mountain diorite) is fine-grained and dark; it resembles the earlier Governor diorite but is much more uniform in composition and has no important differentiation facies developed. Andesite, andesite porphyry, and dacite dikes are related to this intrusion.

It is believed that the first three intrusions were not separated by any great time break and were derived from the same large magma chamber.

The last intrusion is a felsic rhyolite that penetrates the gabbro. Rhyolite and pitchstone dikes, genetically related to the main body, cut the gabbro and the two diorites.

WORKS TO WHICH REFERENCE IS MADE

Burbank, W. S. (1932) *The Western San Juan Mountains,* 16th Intern. Geol. Congr., Guidebook 19, p. 37.

——— (1933) *Vein systems of the Arrastre Basin and regional geologic structure in the Silverton and Telluride quadrangles,* Colo. Sci. Soc., Pr., vol. 13, p. 140.

Cross, W., and Larsen, E. S. (1935) *A brief review of the geology of the San Juan region of southwestern Colorado,* U. S. Geol. Survey, Bull. 843, p. 47, 57-100.

——— and Purington, C. W. (1899) *Telluride folio* (No. 57) U. S. Geol. Survey, 19 pages.

Daly, R. A. (1914) *Igneous rocks and their origin,* p. 26, McGraw-Hill.

Emmons, W. H., and Larsen, E. S. (1923) *Geology and ore deposits of the Creede district, Colorado,* U. S. Geol. Survey, Bull. 718, p. 12.

Gillson, J. L. (1927) *Granodiorites in the Pend Oreille district of northern Idaho,* Jour. Geol., vol. 35, p. 23.

Washington, H. S. (1917) *Chemical analyses of igneous rocks,* U. S. Geol. Survey, Prof. Paper 99, p. 533.

Exploration Department, Shell Oil Company, Shell Building, Houston, Texas.
Manuscript Received by the Secretary of the Society, January 13, 1941.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 52, PP. 721-772, 5 FIGS. MAY 1, 1941

CHANGES IN SEA LEVEL, POSTGLACIAL UPLIFT, AND MOBILITY OF THE EARTH'S INTERIOR

BY BENO GUTENBERG

CONTENTS

	Page
Abstract	721
Introduction	722
Material	722
Causes for slow variations of tide-gauge readings	722
Method of calculation	726
Eustatic changes	728
Postglacial uplift in Fennoscandia	733
Postglacial uplift in North America	739
Postglacial uplift, summary	750
Viscosity and strength of the earth	751
Conclusions	768
Works to which reference is made	770

ILLUSTRATIONS

Figure		Page
1.	Changes in sea level, 10-year averages	730
2.	Changes in sea level in Fennoscandia, yearly mean values	734
3.	Postglacial uplift in Fennoscandia	738
4.	Postglacial uplift in North America	743
5.	Uplift and its rate in Fennoscandia as a function of time	758

ABSTRACT

Record of tide gauges indicate that sea level generally is rising at an average rate of about 10 cm per century. The uplift in Fennoscandia and North America is investigated, and maps showing the rate of uplift are given. A discussion of the new material and historic evidence leave no doubt that the uplift is a consequence of isostatic readjustment of the equilibrium disturbed by the postglacial melting of the ice. The remaining uplift is about 200 meters in Fennoscandia and possibly more in North America, where the present rate of uplift has its maximum of about 2 meters per century in the region of Hudson Bay. Originally, the time needed to reduce the defect in mass to one half under the regions of uplift was less than 10,000 years, but it has been increasing with time and now exceeds 20,000 years.

Theoretical investigations on the plastic flow in the interior of the earth connected with the uplift are critically discussed and extended. The movements affect the whole interior of the earth below the regions of uplift; their amplitudes decrease slowly in the upper 1000 km. If one assumes a strong lithosphere with a thickness of about 70 km and below the asthenosphere with a viscosity of the order of 10^{22} poises, but little or no strength to prohibit plastic flow, there is no disagreement with observations related to isostasy or deep-focus earthquakes. Tectonic processes connected with isostatic anomalies larger than those in the regions of postglacial uplift must be connected with plastic flow at least down to the core. The importance of the effects of small forces acting during long periods is pointed out.

INTRODUCTION

In many regions of the earth continuous changes of the surface are going on. Such movements have been studied either by using precise geodetic measurements or by investigating the changes in the level of the ocean or of lakes relative to fixed points on the shores. Such investigations are not only of much importance in geophysical questions, but they also furnish fundamental data for use in problems of engineering. (*See, e. g.*, Freeman, 1926; Chamber of Commerce of the District of Levis, 1937.) The author, in 1933, investigated such changes of level in North America, but in the interim a large body of data has accumulated calling for revision and extension.

MATERIAL

The most important publication used in this paper is a summary of monthly and yearly mean heights of sea level derived from observations up to the end of 1936, together with information dealing with the way in which the observations have been taken and reduced. It was issued by the Association d'Océanographie (1940) and supplemented by data for 1937 (Association d'Océanographie, 1939). Most countries cooperated in the compilation of these data, without which the fundamental problem of eustatic changes during the last decades could not have been investigated. For a few of those stations which did not report to the international summary, the levels published by Witting (1918) for 1898 to 1912 were used. In addition, the United States Coast and Geodetic Survey kindly furnished data concerning sea level at United States stations for 1938 and 1939. The Hydrographic Service of the Canadian Department of Mines and Resources furnished similar data for Canadian stations in 1938, and, in addition, provided valuable results on the mean level at Churchill, Hudson Bay, as well as information obtained from gauge records on the Great Lakes in Canada. Mean sea levels on the United States shores of the Great Lakes were kindly provided by the United States Lake Survey Office at Detroit, Michigan, and the United States Engineer Office at Duluth, Minnesota.

CAUSES FOR SLOW VARIATIONS OF TIDE-GAUGE READINGS

Publications of tide-gauge readings purport to provide the mean elevation of sea level during a given period at a given point relative to a supposedly fixed mark on the coast. The problem, then, is to calculate the change in sea level and the change in elevation of the coast. In addition the results are influenced by meteorological effects, conditions of the water and the shore, local effects, and errors. The following sources

of true and apparent changes in sea level at a given point have been considered:

(1) Eustatic changes. Due to various causes, such as changes in the configuration of ocean bottoms and coasts, and processes which remove water from the earth's surface or add water to it, the mean sea level changes constantly. The order of magnitude of such variations during periods of geological history can be estimated by investigating contemporary shore lines in widely separated areas. For recent decades, such changes can be detected by comparing tide-gauge readings in various parts of the world.

(2) Effects of meteorological elements on sea level. Changes in air pressure, wind, rain, flow in rivers, local ice conditions, salinity, and temperature of the water are among the causes to be considered. For a more detailed discussion, the reader is referred to Marmer (1931, p. 52), La Cour (1931), LaFond (1939), Woodworth (1927, p. 12), or Johnson (1929), who in addition has pointed out that these effects may be affected by local changes in the shelves near the shore.

Attempts to calculate the effect of some of the meteorological elements on the sea level have been made by La Cour (1931) and by Witting (1918), whose well-known map of the uplift of Scandinavia is based on a discussion of sea-level changes. Variations in the density of sea water and in the level of equilibrium as a consequence of changes of its temperature or salinity have been investigated by LaFond (1939). The total amount of all these effects may be found from comparing geodetic measurements with tide-gauge data. Avers (1931, p. 219) has summarized some of the results. Thus, at St. Augustine, Florida, the mean sea level has been found to be about 24 cm lower than at Galveston. Toward the north along the Atlantic coast the height of mean sea level increases. At Boston, it is about the same as at Galveston, while at Portland, Maine, it is about 7 cm higher. Along the Pacific coast, the calculated differences are even larger; at San Diego the mean sea level has been found to be 40 cm higher than at Galveston, decreasing to 32 cm at San Pedro, California, increasing to 79 cm at Fort Stevens, Oregon, and it is about 65 cm above the Galveston level in the State of Washington. The result, that the mean sea level along the Pacific coast of the United States is about half a meter higher than along the Atlantic coast, "agrees in sign with that shown by the leveling across the Isthmus of Panama, where it was found that the mean sea level on the Pacific coast is approximately 0.2 meter higher than on the Atlantic coast" (Avers, 1931, p. 219). Similar results have been found in other regions. As pointed out by Dawson (1917, p. 15), mean sea level at Quebec is about 45 cm above mean sea

level at New York. Scholaski (1929) found that the mean sea level at Arkhangelsk is 24 cm above, at Odessa 88 cm below, and at Vladivostok 180 cm below the mean sea level at Cronstadt. In Great Britain the values of mean sea level at Newlyn, Harwich, and Felixstowe are in agreement within limits of 2 cm with the values found from geodetic measurements, but mean sea level is 26 cm higher at Dunbar and 35 cm higher at Aberdeen, according to Vignal and Taton (1935). For Norway, the same authors report an altitude of sea level generally increasing toward the north from an assumed zero at Oslo to 16 cm at Bergen, 35 cm at Kjölsdal, and 66 cm at Heimsjö. Bowie (1929) pointed out that in the northern hemisphere all data available indicate a tilting of mean sea level upward with increased latitude. "It seems reasonably certain," he concludes, "that it is due to meteorological conditions such as prevailing directions of winds and differences in barometric pressure with latitude and to the differences in salinity and density of ocean waters at different places." All these data must be considered as approximations only, since the relative elevations of the points mentioned change in course of time.

To reduce the effects under discussion on the values of the calculated sea levels, the author has given greater weight to a few selected stations where a long series of observations is available, so that the undesirable variations are of small influence; shorter series of other stations have been compared with these. By comparing differences between the data of the second group of stations with the contemporary data of the first group, the effect of meteorological elements is greatly reduced.

(3) Effects of the method of observation. Some of the tide-gauge stations have continuous records, at others the mean sea level is calculated from readings by an observer at certain intervals, from once an hour to once a week—at some stations during the day only, at others day and night, at some without reference to the tides, at others only at high and low water, either day and night, or during the day only. The international summary lists 22 different modes of reduction of observations. As long as a station maintains one and the same mode there is no difficulty, although the accuracy of the result depends on the number of observations. However, at many stations the method has been changed from time to time, and sometimes stations change from a few observations at given hours every day to times related to low and high tides, or vice versa. As the rise and fall of the water usually does not correspond with a curve symmetrical with respect to the "mean level," averages taken from high and low will differ from the "mean" calculated from readings made at certain hours. The difference may be positive or negative and depends on the form of the tide curve. (*See*,

for example, Marmer, 1931, p. 51.) In some instances, corrections have been applied before the values were reported to the international summary, for example by the Dutch stations; in others, where no mention is made of such a correction, the data have been used only if no other data near the points in question were available.

At many stations the reference point of the tide gauge has been changed, sometimes repeatedly. As long as the data are in agreement and values before and after the change differ only within the usual fluctuations, this is not serious, although the amount of calculation is increased. However, at many stations, the relationship between the two reference points is not given and in most instances probably is not accurately known. As a consequence, the data before and after the change are not comparable, and valuable information is lost. For this reason, a relatively large number of observations for 1937 could not be used, though some of the missing information may possibly be supplied in later summaries. Where the differences are small, it is often impossible to find out whether the data given in the summary are calculated by taking account of the difference between the reference points. The possibility of mistake is increased by the fact that some stations reckon the sea level from a mark above the average downward, so that higher readings correspond to a lower level, whereas usually the sea-level datum is taken higher with algebraic increase in the figures which, besides, may be positive or negative. The use of various units is an additional source of errors and necessitates calculations.

(4) Effects of changes affecting the tide gauge. At several stations there are sudden jumps in the readings from one year to another without any similar indication at neighboring stations. In such instances it seems likely that the tide gauge or a part of the recording system has moved. In some instances the ground may have settled, in others it may be the effect of an earthquake, either on a district of large area, or on the ground near the position of the gauge, or both. Besides some instances to be given later for several places, *see*, for example, Woodworth's (1927, p. 74) description of effects of the earthquake of 1899 at Yakutat Bay, Alaska, or Tsuboi (1939).

(5) Effects of errors. Errors may occur in reading the gauge, in calculating daily, monthly, or yearly averages, in writing them down, or in printing. So long as they are small, they probably escape attention. However, during the calculations made by the author, discrepancies led to the discovery of a number of misprints in yearly averages, or mistakes in calculating them. In other instances, the cause of suspected errors could not be found.

(6) Effects of land movements. If the tide-gauge readings are corrected for all the effects just mentioned, the resulting figures indicate the change in elevation of the point at which the tide gauge is located. By combining the data for a larger area the gradual changes can be found. The results may be compared with the findings from geodetic measurements, and their reliability may be checked. Unfortunately, both

TABLE 1.—*Uplift in Finland by various methods in cm per century*
The first three lines after Kukkamäki (1939), the last from the present paper.

Station	Turku (Abo)	Hanko	Helsinki	Hamina	Viipuri
Geological observations	74	56	41	31	19
Precise leveling	54	43	36	38	28
Mareographs (Witting)	52	40	36	40	40
Mareographs, present paper	(52)	41	45	(32)	30

types of results are available for a few regions only. The most extensive data exist in Finland, which provides an outstanding example for the conduct of all types of geodetic measurements with important results. Kukkamäki (1939) has compared the findings of the change in elevation from geodetic, geologic, and mareographic investigations (Table 1). The geological results are those of Sauramo (1939), based on the uplift of the shore lines since Littorina I, about 7000 years ago. The geodetic data are from recent measurements, relative to Helsinki for which an uplift of 36 cm per century has been assumed in order to make the data of the second and the third lines equal for this point. The findings based on mareograph records given in the third line are those by Witting (1918). In the last line results of the present paper are added, those in parentheses being interpolated. Kukkamäki's map as well as that of Sauramo have been reproduced by Daly (1940, p. 315, 313, respectively). The agreement between the results of the various methods, especially the three last lines of Table 1, is very good, and the direction of the isobases (lines of equal uplift) drawn from the geodetic data is about the same as from the gauge readings. Thus, the data found from tide-gauge readings extending over a few decades and carefully used are a good substitute for geodetic measurements in investigating tilting of the earth's crust. The geological observations indicate a larger tilt rate as is to be expected, as the rate of tilting decreases with time.

METHOD OF CALCULATION

In using the data from mareographs, two major questions arise: What is the minimum number of yearly means needed to furnish useful results,

and how are they to be combined? Some information on the first question is available in the literature. Marmer (1931, p. 64) points out that 19 years of observation (a full tidal cycle) frequently are considered as necessary to provide a

> "primary determination of mean sea level," but that "a close approximation to mean sea level may be derived from observations covering a period of nine years. . . . Very satisfactory determinations of mean sea level may be derived from observations covering periods of from one to three years, by taking advantage of the fact that the variation in sea level from year to year is much the same over considerable areas. By comparing with simultaneous observations at a station at which long series of observations is at hand, corrections may be derived for reducing the secondary determination of sea level to a mean value." If this method is used, "a year of observation will give a mean sea-level determination correct to within 0.05 foot and four years will give it correct to within 0.02 foot."

In other words, if we have two stations with complete series of observations covering the same 5 years, the average difference between the level at the stations found for this interval is probably accurate within at least ½ cm. The accuracy depends on the local conditions; where the effect of meteorological or other causes is large, the probable errors are greater than in a region with smaller fluctuations.

The previous discussion refers to the problem of finding the mean sea level, but in the present paper we are interested in calculating the changes in level. The first assumption which has been generally made thus far is that the changes are proportional to time. For long periods this is a very good approximation as a comparison of the results of recent geological studies with those provided by geodetic or mareographic information indicates (Table 1). However, the findings leave no doubt that the changes sometimes occur more rapidly, sometimes more slowly. This fact has been recognized by practically everybody working on this problem. Thus, it becomes still more important to use long series of observations at a few places as basic data and to compare the readings at other stations with them.

This leads us to the question of finding the best method for using the data. The most accurate (but also the most tedious) is to use the method of least squares in each case as the author did in his previous paper (Gutenberg, 1933). However, considering the large quantity of data to be digested in the present paper, the author has followed the example of others, and taken the average of the first x years and the average of the last x years, at every station, and divided the difference between the two means by the number of years between the weighted midpoints of these intervals. The problem is to find the best value of x. From the results on accuracy of mean sea level given above, an interval of 10 years for x should be sufficient. This consequently was used in all instances

where the data covered more than 25 years. Where readings for less than 25 years were available, x was taken as one third or slightly more of the whole period. If x is increased, the accuracy of the mean value for each of the two sections is increased, but the time interval between the midpoints of the two periods decreases and thus the error of the result increases. If the number t of observations is large enough to replace the true errors of the observations by the corresponding function of the probable error E of each observation, x should be one third of the whole interval t to make the probable error of the final result a minimum. The probable error of the mean of x observations is $E/\sqrt{x}$, and the probable error of the result $E\sqrt{2}/\sqrt{x}(t-x)$ with a minimum for $x = t/3$.

The procedure for finding the change at a station relative to the change at a near-by station is similar; first the differences between the yearly mean sea levels were calculated, and the results divided in sections as given above, with $x = 10$ for $t > 25$. If two series of observations of 4 or 5 years are available at both stations, covering the same periods, and separated by at least 5 years with or without observations, the comparison, in general, will furnish a result useful for our purpose. Under these circumstances the relative mean sea level at the midpoints of each period then can be found usually within about ½ cm, and the relative change in level within less than 10 cm per century. The results of the present investigation confirm these theoretical predictions, for the large majority of stations. In some instances, however, larger discrepancies indicate relatively large local disturbances or errors.

The discussion, thus far, has dealt with readings of tide gauges at the sea coast. Data on lake levels are to be treated similarly. In this case the effects of wind and air pressure are usually smaller, and the effects of flow of water in rivers, rain, and freezing temperature are larger; however, this affects the whole lake more or less equally so that comparisons between gauges located in the same lake are not much influenced. This is true, too, for the change in the lake level due to the uplift or sinking of the outlet relative to other points of the lake.

EUSTATIC CHANGES

In order to find the uplift of land in a given region by using the change in sea-level, the average change of mean sea level over the whole earth must be known. The sea-level change has been calculated for all stations for which sufficient data are available, omitting, however, those regions where postglacial uplift of the land is known to occur. The results found by employing the methods described in the preceding section are given in Table 2. The data for the stations given in the first column

TABLE 2.—*Changes in sea level in cm per century*

Station	Region	Time (years)	Change	
			station	region
Aberdeen	England	1862/1913	− 1	− 2
Dunbar		1914/1937	− 6	
Liverpool		1857/1937	0	
Le Havre	France (Atlantic)	1860/1886	0	+ 2
Cherbourg		1860/1884	+11	
		1891/1937	0	
Brest		1807/1936	+ 8	
Biarritz		1889/1937	− 9	
Port Vendres	France (Medit.)	1888/1937	+11	+ 9
Sète		1888/1937	− 8	
Port-de-Bouc		1894/1937	+26	
Martigues		1894/1937	+ 3	
Marseilles, P.V.		1890/1926	+22	
Marseilles, M.		1885/1937	+16	
La Ciotat		1893/1926	− 8	
Nice		1888/1909	+11	
Cadiz	Spain	1880/1923	+15	+ 6
Alicante		1874/1934	− 3	
Horta	Azores	1906/1937	+ 9	+ 9
Ajaccio	Corsica Sardinia	1912/1937	+10	+13
Cagliari		1897/1934	+16	
Atlantic City	U.S.A. (Atlantic)	1912/1939	+34	+26
Baltimore		1903/1939	+25	
New York		1843/1902	+23	
		1893/1939	+18	
Balboa	Canal Zone (Pacific)	1909/1934	+13	+13
S. Diego-L. J.	U.S.A. (Pacific)	1906/1939	+17	+12
Los Angeles		1924/1939	+14	
S. Francisco		1898/1939	+12	
Seattle		1899/1939	+ 6	
Honolulu	Hawaii	1905/1939	+22	+22
Sydney	Australia	1897/1927	− 1	− 1
Aden	Arabia	1880/1920	+ 3	+ 3
Bombay, A. B.	India	1878/1936	+ 7	+ 7
Bombay, P. D.		1888/1920	+ 2	
Karāchi		1868/1920	+ 7	
Madras		1880/1920	+11	

Station	Region	Time (years)	Change	
			station	region
Porto Maurizio	Italy (Thyrrh.)	1897/1922	+21	+20
Genova		1884/1936	+13	
Civitavecchia		1896/1921	+ 8	
Napoli, Arsen.		1899/1921	+32	
Napoli, Mandr.		1897/1921	+27	
Palermo	Sicily	1897/1922	+ 6	+ 6
Catania		1897/1919	+ 7	
Porto Corsini	Italy (Adria)	1897/1921	+37	+27
Venezia, Lido		1917/1934	+31	
Venezia, Arsen.		1889/1913	+23	
Venezia, Stef.		1896/1919	+26	
Trieste		1905/1936	+19	
La Goulette	Tunisia	1889/1937	+ 3	+12
Bône	Algeria	1889/1925	+ 9	
Alger		1905/1937	+16	
Oran		1890/1931	+23	
Belgrano	Argentina	1915/1936	+22	+ 9
Mar Del Plata		1911/1937	+ 2	
Buenos Aires		1905/1937	+ 2	
Cristobal	Canal Zone (Atlantic)	1909/1939	− 4	+21
Galveston	U.S.A. (Gulf)	1909/1939	+48	
Key West		1913/1939	+18	
Rangoon	Burma	1880/1920	0	+ 6
Moulmein		1880/1920	0	
Port Blair	Andaman Is.	1880/1920	+16	
Kirun	Formosa	1904/1924	+11	+18
Takao		1904/1933	+25	
Hukabori	Japan	1900/1924	+13	+13
Hamada		1900/1924	+ 8	
Iwasaki		1900/1924	+26	
Otaru		1906/1933	+20	
Hanasaki		1900/1924	+37	
Aikawa		1900/1924	+11	
Hososima		1905/1933	−25	
Average of all stations			+12	
Average of all regions				+11

are combined to form regional averages in order to reduce the effect of the uneven distribution of stations. The third column shows the time interval used.

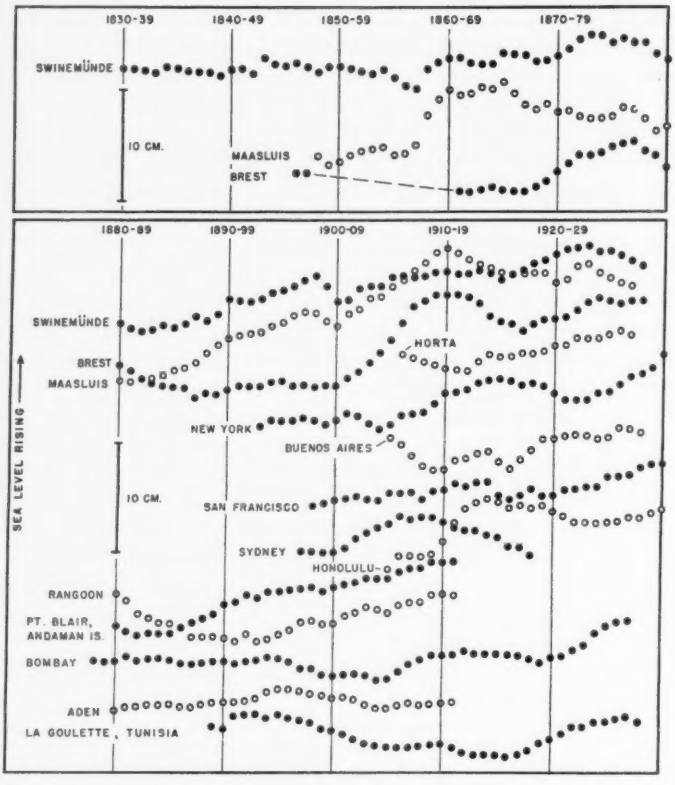


FIGURE 1.—*Changes in sea level, 10-year averages*

The averages of the single stations, as well as of the regional values, indicate that the sea level has risen at a rate of about 11 cm per century during recent decades. This value cannot be influenced much by the settling of tide gauges at some stations. In figure 1[1] 10-year averages have been plotted. They show that the changes are not regular and that the periods of faster rising did not occur simultaneously at all stations.

[1] The figures of this paper were drafted by Mr. John M. Nordquist.

Besides, it seems as if the rising has occurred at a faster rate during recent decades.

The rise of the water level of the ocean may be partly due to an uplift of the bottom in the Hudson Bay, the Gulf of Bothnia, and other oceanic regions with post glacial uplift. However, a noticeable postglacial uplift probably occurs in less than one per cent of the total area occupied by oceans and seems to explain at best 10 per cent of the present rise in ocean level. Sedimentation, too, is a minor cause as has been pointed out for example by Kuenen (1940). Melting of glaciers and of polar ice caps, spreading of continents, uplift of ocean bottoms due to subcrustal currents may contribute to the result. Whatever the cause of the rise in sea level may be, presumably apparent sinking of the land by about 10 cm per century would be found from tide gauges in a stable region. For this reason, 10 cm should be added algebraically to changes in the elevation of land which have been deduced from tide-gauge readings at oceanic coasts during the past few decades.

Certain special remarks about Table 2 are needed. The data for Newlyn, England (50.5°N., 5.5°W.), indicate a rather strong "rise in sea level," about 40 cm per century, but cover only 21 years and have not been included in the table. The data of the Spanish stations are based on incomplete yearly values. The data for the Portuguese stations have been omitted from the table, as the records of Cascais (Lisbon) cover only 21 years, whereas Lagos reported from 1909 to 1937, but with an incomplete series for the first 10 years. Both indicate a fall in sea level at the rate of about 30 cm per century. The findings for most of the Italian stations are partly based on incomplete series of observations. On the other hand, the agreement between the two stations at Napoli and the three at Venezia is good and indicates the accuracy of the method. This is also true for other places with two different series of observation.

At Messina, "in consequence of an earthquake on 28th December 1908 the gauge-zero fell 0.57 m.: in the following years the gauge continued to fall" (Association d'Océanographie Physique, 1940, p. 145). Averages of the "mean sea level" are as follows (in cm):

1897/1901	1903/07	1908 Jan.-Nov.	1908 Nov.	1909 Apr.	1909 May	1909 Apr.-Dec.	1910	1911	1912	1913/17	1918/22
−110	−109	−113	−104	−72	−68	−60	−50	−48	−44	−36	−30

The bench mark is given as at −110 cm until the end of 1908 and at −10 cm thereafter.

There are three stations giving data for the region of the Suez Canal from 1923 to 1937 with some years missing. The results have been

omitted from the table on account of the short time intervals. They indicate a rise of the sea level at Port Said at the rate of about 30 cm per century, of about 5 cm per century at Kabret, and a fall of about 5 cm per century at Port Thewfik.

Some of the stations in Argentina have not been used because of rather incomplete series of observations. The data indicate a rise in sea level by about 20 cm per century at Deseado and by about 40 at Comodoro Rivadavia. The data for Quequén show a rapid decrease of about 20 cm between June and July 1915 and have been omitted.

The data for New York 1843/1902 have been taken from Burr *et. al.* (1904). However, according to information supplied by the United States Coast and Geodetic Survey, the observations prior to 1893 were made for purposes which did not include precise determinations of sea level.

There are some data for New Zealand, but the gaps between them are too long to furnish accurate results. They indicate a slight fall in sea level at Auckland and a small rise at Wellington (in feet):

Auckland: 1900 and 1909.....	6.31	Wellington: 1901 and 1909.....	2.83
1917–1919.....	6.25	1915 and 1919.....	2.95
1921–1923.....	6.23	1921/24, 1927, 1930...	2.93

In Australia, most stations were installed only recently. Sydney has a complete record from 1897 to 1927 and data for several years between 1873 and 1888, as well as for the years 1934 and 1937. There is, however, a sudden change of about ½ foot between 1880 and 1881, so that the older data have not been used in Table 2. Observations at Port Adelaide from 1882 to 1891 give an average sea level of 4.18 feet above the local zero, and for 1933 an average of 4.83 feet. This would indicate a rise of sea level at the rate of about 40 cm per century.

At Tamatave, Madagascar, the average for 1934/37 is about 1 cm higher than the average for 1930/33.

The values for India are of special interest on account of the gravity anomalies. The four stations in Table 2 do not indicate any peculiar effect. However, at Kidderpore (22.5°N., 88.3°E.), there is a gradual decrease in sea level between 1890 and 1900 of about 20 cm, followed by relatively small, irregular changes with a tendency to fall still further.

The changes in Japan have been summarized by Tsuboi (1939) with many references to detailed investigations. The data on mean sea level published by the Association d'Océanographie (1940) contain the effects of the great earthquake of 1923 on the tide gauge at Aburatubo (35.1°N., 139.6°E.). The upheaval at the fixed point at the time of the shock

was estimated as 1.328 meters. During the following years, the water level continued to fall slightly, or the land to rise, as indicated by the following mean yearly sea level (in cm.):

1922	1923	1923	1924	1925	1926	1927	1928	1929
before		after earthquake						
+131	+131	−2	−7	−6	−9	−9	−7	−5

Several other Japanese stations have been omitted from Table 2 because of relatively large irregularities within short periods. As changes in elevation of the land have been found by precise leveling, it was to be expected that in some instances the tide-gauge readings should be affected by them. For details the reader is referred to Tsuboi (1939).

POSTGLACIAL UPLIFT IN FENNOSCANDIA

The fundamental investigation on the uplift now going on in Fennoscandia is set forth in the paper by Witting (1918). His figure showing the lines of equal uplift (isobases) at present is still used in textbooks. (*See,* for example, Daly, 1940, p. 314.) In 1922, he presented a similar figure based on a limited quantity of data from observations extending over more than 100 years; most of these are reproduced in Table 3. In the interim, Egedal (1933) has investigated the uplift in Denmark; his findings are listed by Tsuboi (1939). Other data for northern Scandinavia have been reported by Gavelin (1930) and have been discussed by Tsuboi (1939) (Table 3). The agreement among all these data is good. Considering the fact that longer time intervals are available now, it seemed advisable to use all the information to construct a new map of the uplift in Fennoscandia. The material and the methods have been discussed already. For each station with a record long enough to provide results of sufficient accuracy, the mean uplift has been calculated, taking into consideration the finding that 10 cm need to be added to account for the eustatic change. This correction has also been added to the data used in Table 3 although the figure given is possibly slightly too large for data prior to 1880 (Fig. 1). In addition, the relative changes between pairs of stations were calculated, and the results are found in Table 4. Figure 2 is an example for the remarkable parallelism between the yearly changes from station to station in the same region.

In Table 4 the first column shows the number which was given to the stations; the third column, the time interval between the first and the last reported yearly mean sea level. In the sixth column, the uplift is calculated from x years at the beginning and end of the series following the rules discussed in the section on methods; 10 cm have been

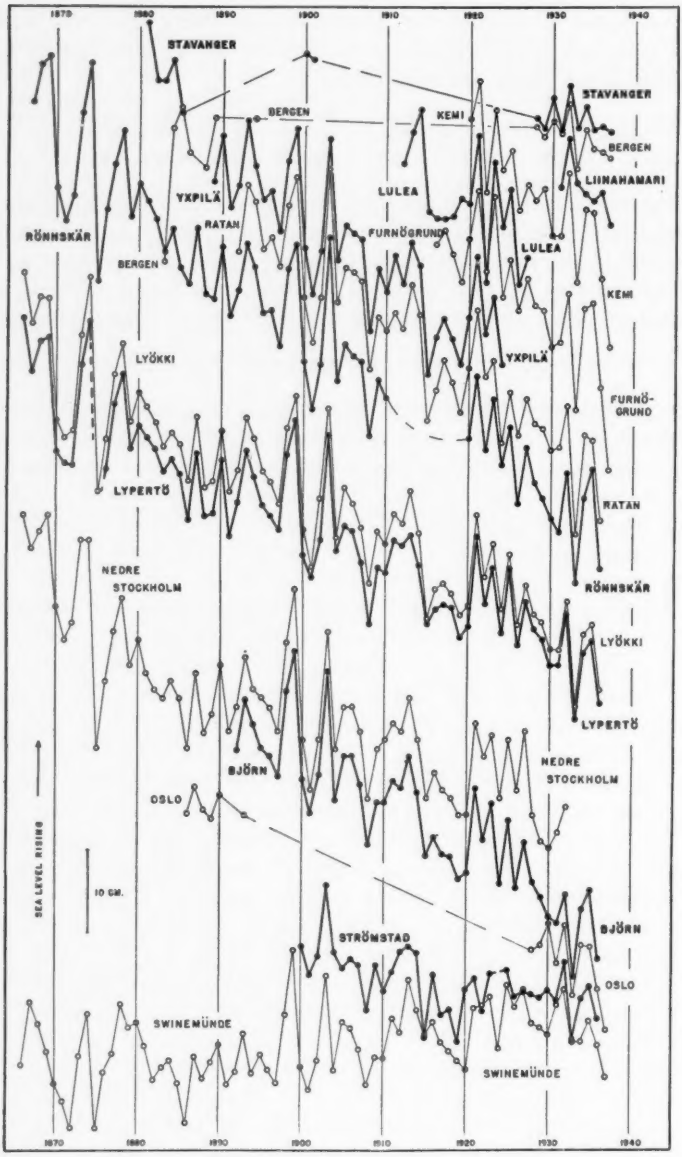


FIGURE 2.—*Changes in sea level in Fennoscandia, yearly mean values*

added to the result. In the following three columns the relative changes between pairs of stations are listed; in each column, the first figure indicates the number of the second station used; the second figure, the difference in the rate of uplift in cm per century; and the third

TABLE 3.—*Uplift of land in Fennoscandia in cm per century*
First section after Witting (1922), second section after Tsuboi (1939). 10 cm are added. (*See* text.)

No.	Station	Time Years	Lat. N.	Long. E.	Uplift
84	Skallö (Karlskrona)	1755–1888	56.7	16.4	24
85	Landsort	1800–1905	58.8	17.9	53
86	Lövgrund	1731–1869	60.8	17.5	85
87	Ledskär (Ratan)	1749–1905	64.0	20.9	107
88	Rataskär (Ratan)	1774–1905	64.0	20.9	105
89	Storrebben (Piteå)	1648–1750	65.2	22.0	110
90	Wattungen	1697–1920	63.1	20.8	105
91	Bergö	1755–1920	63.0	21.2	110
92	Utö	1800–1905	59.8	21.4	47
93	Hangö	1754–1920	59.8	22.9	48
94	Jusarö	1800–1920	59.8	23.6	55
95	Löparö	1750–1920	60.2	25.5	30
96	Lovisa	1759–1920	60.4	26.3	28
97	Kronstadt	1841–1913	60.0	29.9	0?
98	Piteå district	1654–1908	65	22	114
		1654–1750			126
		1750–1851			104
		1785–1884			101
		1796–1908			97
		1869–1908			89

figure, the uplift at the station under consideration under the supposition that the uplift given in the last column for the second station is correct. Of course, this procedure must start at one or more stations with long series of observations. In Table 4, it was assumed that the uplift found from the combined data for Stockholm and Nedre Stockholm (112 years) and those for Swinemünde (126 years) are correct. Then for the neighboring stations the differences with respect to the basic stations were calculated; weighted averages between the values in column 7 and those found from the tide-gauge readings at the station (column 6) were entered in the last column. This procedure then was followed for the other stations, using sometimes as many as three different combinations. In general, the agreement between the values found in different ways is very good, indicating the degree of accuracy

of the method. Some exceptions or noteworthy facts will be discussed next.

Comparison of the mean sea level at Toppila with the values at other stations shows a sudden rise of about 10 cm between 1895 and 1896. As this seems to be due to a change in the zero point of the gauge, the values for the years 1889 to 1895 have not been used. Many of the following stations in Table 3 are given with various names, for example, Yxpilä = Ykspihlaja, or Alholmen = Leppäluoto = Pietarsaari. In this paper, only one name is used. The co-ordinates (columns 3 and 4 in Table 4) are therefore added.

For many of the stations in the Gulf of Bothnia, the data of the stations at Lyökki and Lypertö were used for comparison. Both have records of 78 years with data rarely incomplete, and their yearly mean sea level agrees excellently. The following are differences in mean sea level in cm from 1860 to 1869: 4.4, 4.3, 4.0, 4.0, 2.8, 3.3, 3.3, 3.7, 3.0, 2.5; the differences continue to decrease gradually, and from 1927 to 1936 they are: 0.0, 0.1, 0.0, 0.0, —0.3, —0.3, —0.4, 0.2, 0.1, —0.4.

The data for Hogland, Reval, and Takhona were taken from Witting (1918). For the first two some more recent observations are available, but the connection of the zero points for the two sets of data is doubtful.

For the Danish stations, the agreement between the findings of Egedal (1933) and the new values is good. The following are the values for the rate of change in mean sea level in cm per century for various stations (numbers as in Table 4), *a* after Egedal, *b* calculated in this paper:

No.	51	52	53	54	55	56	57	58	59	69
a	5	6½	2½	12½	3	13½	8½	3½	2	16½
b	4	2	1	10	3	9	11	1	−1	7

Note that the values in Table 4 are for uplift of land, with 10 cm added for eustatic changes.

Unfortunately, the data for Norway had to be based on short, incomplete series. On account of their importance for the construction of isobases in the western half of the region under consideration, they are discussed here more fully. For Oslo, more or less complete series of observations are available for 1886/90, 1892/3, and 1928/37. The following is the resulting mean sea level (in cm):

1886/90	1892/3	1928/32	1933/37
+12.8	+10.6	−4.5	−9.3

From these data, as well as comparison with other stations (Table 4; Fig. 2), it may be safely concluded that the uplift of the land there is

TABLE 4.—*Uplift of land in northwestern Europe*

No.	Station	Time Years	Lat. N.	Long. E.	Uplift in cm per century									Assumed
					Dir.	From comparison with other stations (no., difference, result)								
1	Kemi	1920/1937	65.7	24.5	125	2	+13	101	3	0	105	10	+35	102	105
2	Toppila	1896/1937	65.0	25.4	93	12	−14	89	10	+17	84	4	+ 7	88	88
3	Hornankallio	1923/1937	64.7	24.5	125	12	0	103	2	+14	102	.....			105
4	Yxpilä	1889/1924	63.8	23.0	81	10	+13	80	6	− 7	83	.....			81
5	Alholmen	1915/1937	63.7	22.7	108	10	+29	96	.....			.....			100
6	Vaasa-Vasklot	1884/1937	63.1	21.6	87	10	+24	91	7	+10	93	.....			90
7	Rönnskär	1867/1936	63.1	20.8	81	10	+19	86	.....			.....			83
8	Luleå	1912/1927	65.5	22.2	90	13	0	81	6	+19	109	10	+35	102	95
9	Mäntyluoto	1912/1937	61.6	21.5	83	10	+10	77	.....			.....			80
10	Lyökki	1858/1936	60.9	21.2	67	11	+ 5	67	.....			.....			67
11	Lypertö	1858/1936	60.6	21.2	61	17	+12	62	.....			.....			62
12	Furnögrund	1916/1936	64.9	21.2	101	13	+18	99	5	+ 7	107	.....			103
13	Ratan	1892/1936	64.0	20.9	81	10	+10	77	4	+ 3	84	.....			81
14	Draghällan	1898/1936	62.3	17.5	82	10	+20	87	17	+36	86	13	+ 5	86	85
15	Nedre Gävle	1896/1936	60.7	17.2	64	11	+ 5	67	14	−22	63	.....			65
16	Björn	1892/1936	60.6	18.0	66	17	+17	67	11	+ 5	67	.....			67
17	Stockholm	1889/1936	59.3	18.1	45	18	0	50	.....			.....			50
18	Ned. Stockholm	1825/1927	59.3	18.1	51	.....			.....			.....			50
19	Ned. Södertälje	1869/1936	59.2	17.6	44	18	−10	40	.....			.....			42
20	Strömma	1913/1936	60.2	22.9	58	11	− 4	58	.....			.....			58
21	Jungfrusund	1858/1934	60.0	22.4	40	11	−19	43	.....			.....			41
22	Utö	1866/1936	59.8	21.4	37	11	−22	40	21	− 7	34	.....			37
23	Hangö	1866/1936	59.8	23.0	41	22	+ 4	41	.....			.....			41
24	Helsinki	1879/1937	60.1	25.0	44	23	+ 6	47	.....			.....			45
25	Söderskär	1866/1936	60.1	25.4	33	22	− 4	33	.....			.....			33
26	Viborg	1889/1937	60.7	28.7	27	25	− 2	31	.....			.....			30
27	Hogland	1898/1912	60.1	27.0	32	26	−21	9	37	+29	39	.....			25
28	Reval	1898/1912	59.5	24.5	33	24	−31	14	37	+ 6	16	.....			20
29	Takhona	1898/1912	59.1	22.6	19	21	−21	20	37	+25	35	.....			25
30	Landsort	1887/1936	58.8	17.9	35	17	−12	38	.....			.....			37
31	Mem	1864/1924	58.5	16.4	30	18	−24	26	.....			.....			28
32	Västervik	1896/1927	57.8	16.6	18	30	−19	18	34	− 1	11	.....			16
33	Kalmar	1903/1927	56.7	16.4	37	32	+19	35	34	+15	27	.....			33
34	Karlskrona	1887/1936	56.1	15.6	10	17	−36	14	30	−24	13	.....			12
35	Ystad	1887/1936	55.4	13.8	− 4	34	−15	−3	45	− 2	3	.....			0
36	Malmö	1902/1927	55.6	13.0	−12	35	− 4	−4	52	− 8	0	.....			− 4
37	Dünamünde	1872/1936	57.1	24.0	10	34	+ 7	19	19	−35	7	35	+18	18	10
38	Riga	1872/1936	57.0	24.1	9	37	+ 5	15	.....			.....			10
39	Domesnäs	1884/1931	57.8	22.6	− 4	38	−11	−1	41	− 4	1	.....			− 1
40	Windau	1873/1936	57.4	21.5	0	38	− 1	9	.....			.....			4

31	Memel	1898/1927	55.5	13.4	36	43	23	26											25
32	Västervik	1896/1927	57.8	16.6	18	30	−19	18	34	− 1	11								16
33	Kalmar	1903/1927	56.7	16.4	37	32	+19	35	34	+15	27								33
34	Karlskrona	1887/1936	56.1	15.6	10	17	−36	14	30	−24	13								12
35	Ystad	1887/1936	55.4	13.8	− 4	34	−15	−3	45	− 2	3								0
36	Malmö	1902/1927	55.6	13.0	−12	35	− 4	−4	52	− 8	0								− 4
37	Dünamünde	1872/1936	57.1	24.0	10	34	+ 7	19	19	−35	7	35	+18	18					10
38	Riga	1872/1936	57.0	24.1	9	37	+ 5	15											10
39	Domesnäs	1884/1931	57.8	22.6	− 4	38	−11	−1	41	− 4	1								− 1
40	Windau	1873/1936	57.4	21.5	0	38	− 1	9											4
41	Libau	1865/1937	56.5	21.0	9	38	− 9	1	45	− 3	2	34	− 2	10					5
42	Memel	1898/1919	55.7	21.1	18	45	+20	25	43	+ 5	10								18
43	Pillau	1898/1937	54.6	19.9	6	45	0	5											5
44	Stolpmünde	1911/1937	54.6	16.9	14	45	− 2	3											9
45	Swinemünde	1811/1937	53.9	14.3	5														5
46	Arkona	1882/1936	54.7	13.4	4	45	+ 9	14	47	+14	9								9
47	Warnemünde	1882/1937	54.2	12.1	−14	35	− 3	− 3	48	0	− 2	45	− 6	− 1					− 5
48	Wismar	1849/1881	53.9	11.5	− 2	45	+ 3	8											
		1882/1937			−14	45	− 6	− 1											− 2
49	Travemünde	1882/1937	54.0	10.9	−14	48	− 2	− 4											− 8
50	Marienleuchte	1882/1937	54.5	11.3	− 4	48	+ 7	5											0
51	Gedser	1898/1937	54.6	12.0	8	52	− 2	6											6
52	Köbenhavn	1889/1937	55.7	12.6	7	48	+15	13	35	+ 8	8	34	− 7	5					8
53	Hornbaek	1898/1937	56.1	12.5	12	60	− 4	7											9
54	Fredericia	1890/1937	55.6	9.8	1	52	− 8	0											0
55	Korsör	1897/1937	55.3	11.1	9	50	+ 4	4	51	+ 2	8								7
56	Slipshavn	1896/1936	55.3	10.8	3	55	− 5	2	50	− 2	− 2								1
57	Aarhus	1889/1937	56.1	10.2	2	59	−15	− 4											− 1
58	Frederikshavn	1893/1937	57.4	10.6	13	61	−17	7	59	− 2	9								9
59	Hirtshals	1892/1937	57.6	10.0	15	54	+14	14	48	+ 9	7	60	+ 1	12					11
60	Varberg	1887/1936	57.1	12.2	7	52	+ 8	16											11
61	Göteborg	1887/1936	57.7	12.0	23	52	+23	31	60	+10	21								24
62	Bäckevik	1895/1928	58.4	11.3	26	61	0	24	59	+24	35								28
63	Strömstad	1900/1936	59.0	11.2	28	61	+15	30	59	+18	29								32
64	Oslo	1885/1937	59.9	10.7	57	60	+38	49	48	+65	63	61	+40	64					58
65	Stavanger	1881/1937	59.0	5.7	18	71	+11	6											12
66	Bergen	1883/1937	60.4	5.3	9	71	+27	22											17
67	Narvik	1928/1937	68.4	17.4	?	66	−30?	−13?											−15?
68	Liinahamari	1931/1937	69.6	31.4	?	66	+30?	47?	67	+55?	40?								45?
69	Esbjerg	1889/1937	55.5	8.5	2	71	+ 9	4											3
70	Bremerhaven	1898/1937	53.5	8.6	3	71	+19	14	69	+ 5	8								8
71	Delfzijl	1865/1937	53.3	6.9	0	59	−24	−13											− 5
72	Harlingen	1865/1936	53.2	5.4	− 8	71	− 7	−12											−10
73	Den Helder	1865/1937	53.0	4.8	− 2	76	− 7	−11											− 8
74	Ijmuiden	1871/1937	52.5	4.6	12	73	+17	9											10
75	Hoek v. Holland	1864/1937	52.0	4.1	−17	76	−19	−23											−19
76	Maassluis	1848/1936	51.9	4.3	− 4	71	+ 2	− 3											− 4
77	Hellevoetsluis	1861/1937	51.8	4.1	− 4	76	− 2	− 6											− 5
78	Brouwershaven	1872/1937	51.7	3.9	4	77	+ 4	− 1											0
79	Zierikzee	1872/1937	51.6	3.9	− 2	74	− 6	− 6											− 4
80	Vlissingen	1862/1936	51.5	3.6	11	77	+13	8											9
				West															
81	Liverpool	1857/1937	53.4	3.0	10														10
82	Dunbar	1914/1937	56.0	2.5	4	71	+ 8	3											4
83	Aberdeen	1862/1913	57.1	2.1	11	71	+ 5	0											5

81	Liverpool	1857/1937	53.4	3.0	10	.....	.....	.....		10
82	Dunbar	1914/1937	56.0	2.5	4	71 + 8	3	.....	.....	4
83	Aberdeen	1862/1913	57.1	2.1	11	71 + 5	0	.....	.....	5

at the rate of between 40 and 70 cm per century during recent decades.

The mean sea level at Stavanger is found as follows:

1881/85	1899/1901	1928/32	1933/37
13.1	14.8	8.1	6.6

The data are very incomplete; between 1881 and 1885 there is no year with all 12 monthly averages available. However, a slight uplift of the land can safely be concluded. The reader is referred to Table 4 and Figure 2 for this and the following stations. For Bergen the mean sea level is based on almost complete yearly series for 1883/86, 1888/89, 1893/4, and 1928/37 with the following values (in cm):

1883/6	1888/89	1893/94	1933/37
4.7	4.6	8.3	3.5

These figures clearly indicate no definite change in sea level, and probably a slight uplift of the land not exceeding 20 cm per century, considering the eustatic change. On the other hand, very incomplete data for Trondhjem between 1872 and 1881 suggest a rapid uplift of the land there but are too scanty for definite results: water level 1872/74, 30 cm; 1877/78, 24 cm; 1880/81, 21 cm above an unknown zero.

The data for Narvik cover the years from 1928 to 1937 and are based on incomplete series. The following results are found (in cm):

Sea level Narvik				Difference sea level Narvik-Bergen			
1928/30	1932/33	1934/35	1936/37	1929/30	1932/33	1934/35	1936/37
6	20	16	18	4	11	12½	16

The data, if correct, indicate a slight sinking of the land at Narvik. At Vardö, the mean sea level between 1880 and 1885 (data very incomplete) was about 4 cm lower than in 1891/92, indicating no large change and a sinking of the land rather than an uplift. For other stations in Norway, the observations are insufficient to give even the direction of the change. However, there are readings available for Liinahamari in northern Finland for 1931 to 1937. In general such a short period was not used, but even the order of magnitude of the change in this region is of interest; the following figures are available (in cm):

Sea level at Liinahamari			Difference Bergen—Liinahamari		
1931/33	1934/35	1936/37	1931/33	1934/35	1936/37
8	5	1½	6½	8½	9

The method of least squares indicates a fall of sea level at the rate of about 30 cm per century. All data available point to a rise of the land at the rate of between 20 and 60 cm per century. This agrees

with the findings of Tanner (1930) who concluded from geological evidence that there is probably an uplift at the rate of between 25 and 35 cm per century near the cost of the Varangerfjord plus the effect of a

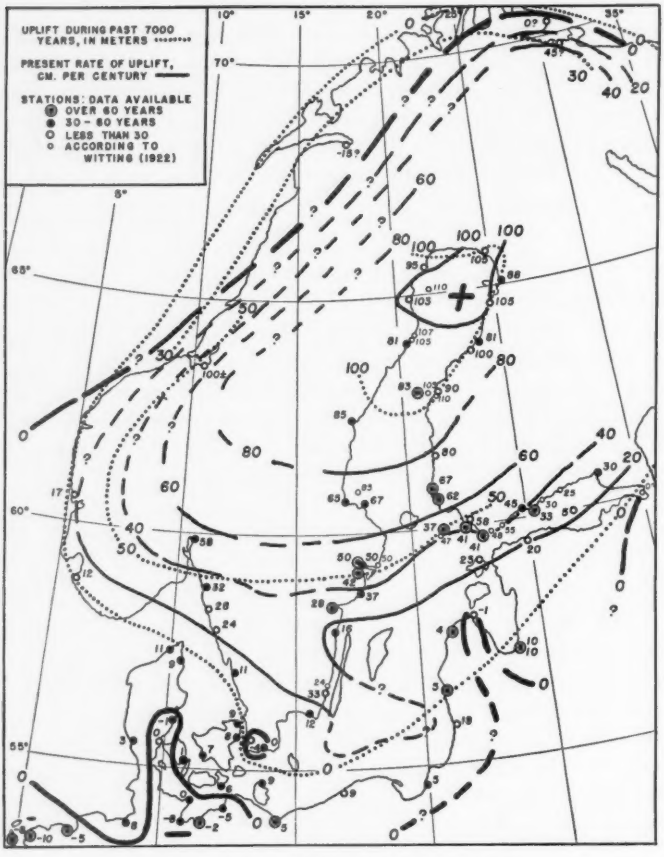


FIGURE 3.—*Postglacial uplift in Fennoscandia*
Data for past 7000 years after Sauramo (1939).

possible eustatic change with isobases bending rather sharply southward if one proceeds toward the east, as indicated in Figure 3. The zero isobase seems to run in a north-south direction near the eastern tip of the Kola peninsula.

The small values for the change in sea level at the Dutch and French stations (Table 4, nos. 71 to 80; Table 2, second section) leave no doubt that these coasts have been very stable during the past 100 years and disprove the conclusion of Schmidt (1922) based on comparison of geodetic surveys in 1857 and 1884, that during this time interval the French coast had been sinking by a maximum of about 1 meter near the Dutch border, by about 80 cm near Brest, and less toward the south. Apparently the precision of the surveys was not high enough to find the correct changes. The same author (Schmidt, 1918) concluded also from geodetic measurements that the northern foreland of the Alps in Bavaria was moving horizontally, but Schwinner (1933; 1935) and others have shown these conclusions to be incorrect and that the hypothetical horizontal movements in this region are within the limits of error of the geodetic data—smaller than at the rate of ½ cm per year.

The results which have been discussed and the data of Tables 3 and 4 were used to construct a new map of the contemporary uplift in Fennoscandia and the surrounding regions (Fig. 3). Isobases are drawn for every 20 cm per century. In the southern part, the zero isobase probably follows the coast of the Baltic Sea slightly to the south. This result agrees with findings of Meissner (1924; 1926), who came to the conclusion after detailed investigations that the coast region of the Baltic Sea in Germany is stable. In spite of very scanty data in the northwestern half of the region, and the possibility that the isobases are closer together in Norway, as indicated by geological data, the general picture is well brought out. The isobases are evidently far apart in the central region. This agrees with the geological findings as to the uplift during the past 7000 years (Initial—Littorina Sea stage). Some of the isobases for this time interval have been added to the figure after Sauramo (1939). As his curves represent the uplift in meters during 7000 years, and the new curves the present rate in cm per century (or meters in 10,000 years), a comparison shows that the recent rate of uplift is about two thirds of the average rate during the past 7000 years. The slowing down of the uplift may also be seen from the last figures in Table 3.

Using data given by various authors, Niskanen (1939, p. 19) calculated the uplift at many places in southern Finland at various distances d from Angermanland. A few characteristic results are given in Table 5.

POSTGLACIAL UPLIFT IN NORTH AMERICA

The findings on postglacial uplift in North America during recent decades have been based thus far only on data furnished by the gauges

in the Great Lakes. In a previous paper the author (Gutenberg, 1933) has discussed the material up to and including 1931. In the interim, readings for the following 8 years have been made available, and some important revisions have been made. In calculating the changes the method of least squares used in the earlier paper was replaced by the

TABLE 5.—*Uplift (in meters) in Finland*
Relative to the present conditions after Niskanen (1939).

d km	Station	7250 B.C.	6800 B.C.	6500 B.C.	5000 B.C.	4500 B.C.	3300 B.C.	1800 B.C.	1000 B.C.	0
220	Lauhavuori	212	192	145	91	80	61	38	28	19
285	Vatula	180	160	126	80	71	55	35	26	17
240	Tampere	143	123	105	68	60	49	33	23	15
395	Hämeenlinna	120	100	94	60	54	45	30	22	14
440	Lahti	110	88	85	56	50	42	28	21	13
475	Helsinki	67	52	47	29	28	28	19	14	9
555	Suursaari	54	43	39	20	22	23	17	13	8
640	Ino	25	21	16	2	8	10	10	8	6

method described in the preceding section, and the relative changes from station to station were not found graphically, but by numerical calculation (Table 7).

The investigation needs to be carried out for each lake independently. Differences between the gauge readings at pairs of stations were calculated; Table 6 gives some examples. In general, the differences change gradually and leave no doubt as to how one point of a given pair has moved relative to the other. Absolute values (corresponding to the column 6 in Table 4) were not calculated, as they are influenced by the general change in the lake level due to precipitation, flow in rivers, vertical movements of the region at the outlet, wind, and other causes. Table 7 gives relative changes between pairs of stations in a way similar to that in Table 4. In the next to the last column the assumed values are listed; an arbitrary zero has been taken for each lake. The results were plotted, and lines of equal relative uplift were drawn for each lake. From the fact that there is practically no relative change between Calumet Harbor and Milwaukee on Lake Michigan, and between Port Stanley and Cleveland on Lake Erie, it is concluded that the zero isobase runs to the north of these stations, and consequently that the zero assumed for these two lakes approximates the absolute zero. This agrees with the findings of Taylor (1926) that, at least since the time of the Nipis-

sing Great Lakes, which was about 5000 years ago according to Antevs (1939), there has been no noticeable uplift in this region. The zero isobase, "Algonquin-Nipissing Hinge line" of Taylor (1926), may be found in Figure 4, according to Daly (1940, p. 327; *see also* Daly, 1934).

TABLE 6.—*Relative changes in lake level in the Great Lakes*

Five-year averages of differences; unit 1/100 foot; zero arbitrary and different in each column.

Years	Duluth-Port Arthur	Duluth-Marquette	Duluth-Michipicoten H.	Marquette-Port Arthur	Calumet Harbor-Milwaukee	Milwaukee-Mackinaw City	Milwaukee-Harbor Beach	Cleveland-Port Colborne	Cleveland-Buffalo	Toronto-Pt. Dalhousie	Toronto-Oswego	Toronto-Kingston
1865/1869	.....	.....	.....	.....	.....	.....	.....	−23	.....	−25	−55	.....
1870/1874	.....	.....	.....	.....	.....	.....	.....	−15	.....	− 3	−42	.....
1875/1879	.....	.....	.....	.....	.....	.....	−19	−15	.....	+14	−40	.....
1880/1884	.....	.....	.....	.....	.....	.....	−23	− 6	.....	+10	−42	.....
1885/1889	.....	.....	.....	.....	.....	.....	−22	−13	.....	− 1	−34	.....
1890/1894	.....	−26	.....	.....	.....	.....	− 7	− 8	−23	−18	−46	.....
1895/1899	.....	−23	.....	.....	.....	.....	−13	− 5	−17	−14	−40	.....
1900/1904	.....	−21	.....	.....	.....	−21	−12	− 5	−16	− 9	−33	−18
1905/1909	.....	−14	.....	.....	.....	−16	−10	− 1	−17	− 2	−22	− 8
1910/1914	.....	−16	.....	.....	0	−16	−11	0	− 9	?	?	?
1915/1919	−31	−14	.....	−17	+ 2	− 8	− 5	+15	− 8	− 4	−14	− 5
1920/1924	−24	−10	−29	−14	− 3	− 7	− 1	+15	− 5	+ 1	− 9	+ 1
1925/1929	−16	−10	−23	− 6	− 7	− 5	− 2	+ 7	− 7	0	−10	+ 1
1930/1934	−12	− 6	−20	− 6	− 4	− 4	− 2	+ 4	− 6	0	− 4	+ 9
1935/1939	− 8	− 3	−13	− 5	− 2	+ 4	0	+ 8	− 3	+ 1	+ 1	+14

The selection of the absolute zero for the data concerning Lake Ontario is not difficult as Port Dalhousie and Port Colborne are not far apart. For Lake Superior to secure accuracy in the assumed absolute zero is more difficult, as the stations are farther apart and the hinge line seems to curve somewhat in this region; however, the error probably is less than 10 cm. The last column of Table 6 was found by using these relationships between the adopted arbitrary zero for each lake and its absolute values. It gives the absolute uplift of the stations in cm per century with errors probably not in excess of 10 cm for most stations.

Brief remarks are needed concerning some of the stations. The main difference between the results of Table 6 and those of the previous paper by the author (Gutenberg, 1933) concerns Calumet Harbor. In the earlier paper, the author pointed out that "the large tilt between Calumet and Milwaukee has looked suspicious to all investigators, but that it has been determined on several occasions that the gauge at Calumet Harbor has been stable." The new data furnished by the United States Lake Survey Office at Detroit show that this suspicion was well founded. Lieu-

TABLE 7.—*Uplift of land in the Great Lakes region*

No.	Station	Time, years	Uplift in cm per century, from comparison with other stations (no., difference, result)			Assumed	
						Relat.	Abs.
101	Calumet Harbor	1904/1939	102 + 3 3	.....	.....	3	3
102	Milwaukee	1860/1939	.....	.....	.....	0	0
103	Mackinaw City	1900/1939	102 +20 20	101 +16 19	.....	20	20
104	Harbor Beach	1875/1939	101 + 8 11	102 +11 11	103 − 9 11	11	11
105	Goderich	1910/1939	102 + 9 9	104 + 5 16	103 −14 6	10	10
106	Collingwood	1906/1939	104 +19 30	105 +15 25	.....	28	28
107	Marquette	1860/1939	.....	.....	.....	0	24
108	Port Arthur	1907/1939	107 +20 20	109 +35 23	.....	21	45
109	Duluth	1888/1939	107 −12 −12	.....	.....	−12	12
110	Michipicoten H.	1915/1939	107 +15 15	108 − 2 19	109 +29 17	17	41
111	Port Stanley	1926/1939	112 0 0	.....	.....	0	0
112	Cleveland	1860/1939	.....	.....	.....	0	0
113	Port Colborne	1860/1939	111 +17 17	112 +16 16	.....	16	16
114	Buffalo	1888/1939	112 +11 11	113 0 16	.....	13	13
115	Pt. Dalhousie	1860/1939	.....	.....	.....	0	20
116	Toronto	1861/1939	115 (0) 0	.....	.....	0	20
117	Oswego	1860/1939	115 (+18) 18	116 +24 24	.....	22	42
118	Kingston	1895/1939	115 +16 16	116 +19 19	117 − 1 21	19	39
119	Cape Vincent	1916/1939	116 +23 23	118 − 4 15	.....	19	39

tenant Colonel Richards has kindly pointed out to the author that the gauge at Calumet Harbor was not a Lake Survey gauge before 1935. When it was taken over, the Engineer Office of the Lake Survey made an extensive study of the records and found that between 1902 and 1935

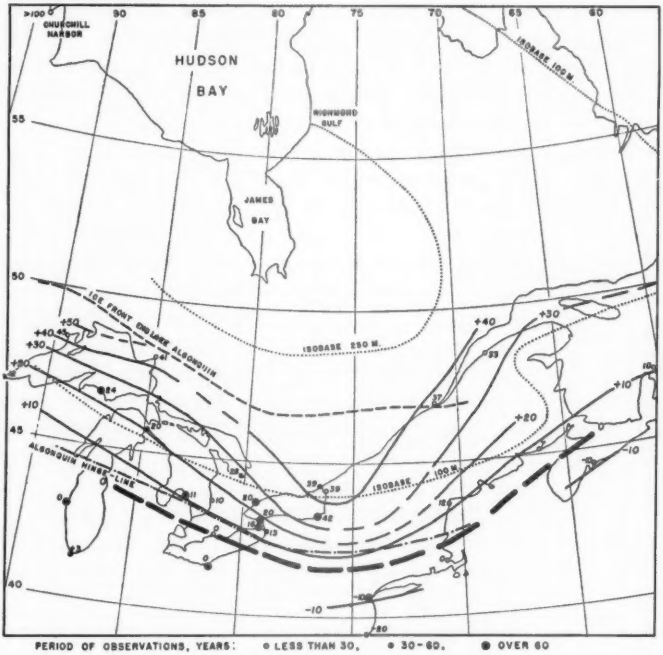


FIGURE 4.—*Postglacial uplift in North America*
Curves based on geological evidence after Daly (1934).

the reference point had settled 0.26 feet with respect to a permanent bench mark well back from the water front. There had been other errors, besides. The new data are considered the best which can be calculated by using the original notes.

For other stations, the changes are much smaller and are noticeable only for Harbor Beach. The data for Toronto for 1911 and 1912 show a deviation of about 0.2 feet against the other stations at Lake Ontario and were not used. The differences between the mean values for Point Dalhousie and Oswego are very irregular for the years preceding 1890 and were omitted. The difference between the data for Toronto and

TABLE 8.—*Uplift (+), or sinking (−) of land in North America*

No.	Station	Time, years	Lat. N.	Long. W.	Uplift in cm per century				Assumed
					Dir.	by comparison with other stations (no., difference, result)			
121	Cristobal	1909/1934	9.3	79.9	+14	122 +51 +13	.....	.....	+14
122	Galveston	1909/1939	29.3	94.8	−38	124 −16 −36	129 −28 −38	.....	−38
123	Pensacola	1924/1939	30.4	87.2	?	122 +29 − 9	124 0 −10	.....	− 9
124	Key West	1913/1939	24.6	81.8	− 8	127 + 4 − 8	129 − 2 −12	.....	−10
125	Fernandina	1898/1923	30.7	81.5	+22	129 +35 +25	.....	.....	+24
126	Charleston	1922/1939	32.8	79.9	?	127 +14 + 2	124 + 8 − 2	129 +11 + 1	0
127	Baltimore	1903/1939	39.3	76.6	−15	129 − 1 −11	.....	.....	−12
128	Atlantic City	1912/1939	39.7	74.4	−24	127 − 5 −17	129 − 9 −19	.....	−20
129	New York	1893/1939	40.7	74.0	−10	.....	.....	.....	−10
130	Boston	1922/1939	42.3	71.1	?	129 +11 + 1	131 −10 − 2	.....	0
131	Portland, Maine	1912/1939	43.7	70.2	+12	129 +21 +11	.....	.....	+12
132	Halifax	1895/1938	44.7	63.6	−11	129 0 −10	131 −16 − 4	.....	−10
133	St. John, N. B	1894/1938	44.3	66.1	+ 9	132 +13 + 3	.....	.....	+ 7
134	St. Paul Island	1895/1926	47.2	60.1	+ 5	135 −19 +14	.....	.....	+10
135	Father Point	1897/1928	48.5	68.5	+35	133 +19 +26	129 +45 35	.....	+33
136	Quebec	1894/1920	46.8	71.2	+38	135 0 +33	129 +40 +30	133 +40 +47	+37
137	Balboa	1909/1934	9.0	79.6	− 3	.....	.....	.....	− 3
138	San Diego-La Jolla	1906/1939	32.8	117.2	− 7	139 − 5 − 7	.....	.....	− 7
139	San Francisco	1898/1939	37.8	122.5	− 2	129 + 1 − 9	.....	.....	− 2
140	Seattle	1899/1939	47.6	122.3	+ 4	139 + 8 + 6	.....	.....	+ 5
141	Victoria	1909/1936	48.4	123.4	+10	140 + 1 + 6	.....	.....	+ 7
142	Vancouver	1902/1935	49.3	123.1	+10	140 − 4? + 1?	.....	.....	+ 7
143	Port Simpson	1902/1920	54.6	130.4	+20?	142 +21? +28?	.....	.....	+24?
144	Ketchikan, Alaska	1919/1939	55.3	131.6	− 1	140 + 8 +13	.....	.....	+ 8

Oswego is not quite so irregular (Table 6) for this period; the difference between the level at Toronto and Port Dalhousie shows relatively large fluctuations about zero (Table 6). Apparently, there are meteorological conditions which affect the mean lake level at the various stations on Lake Ontario more than those in the other lakes and therefore it is to be expected that the results for Lake Ontario are less accurate.

In order to investigate variations in course of time, maps of uplift have been prepared (not published) similar to Figure 4 but based on successive 5-year averages. Before 1900, insufficient data are available, and the few resulting changes seem to be more irregular than in later years. About 1900, the tilt seems to have been small in the Lake Erie region. The map for 1905 based on a comparison of the average differences 1900 to 1904 with those 1905 to 1909 shows tilt above average. The map of 1910 indicates very small tilt. In the Lake Michigan-Lake Huron region, the maximum change was at the rate of 6 cm per century (Harbor Beach-Milwaukee), and its direction was reversed. The figures for Mackinaw City indicate no change beyond 3 cm per century relative to the other stations. In the Lake Superior region, only the pair Duluth-Marquette is available, and it shows a reversed tilt at the rate of 12 cm per century. About 1915, the tilting was very strong. The isobase with an uplift at the rate of 50 cm per century runs approximately through Mackinaw City and Toronto. In 1920 the tilt was slightly less than normal—about 1925 it was normal in the Lake Superior region, small in the northern part of Lake Michigan, across Lake Huron and Lake Ontario, and reversed in the southern part of Lake Michigan and across Lake Erie. For the following period (1930) the picture is almost reversed, small changes had occurred in the region of Lake Superior, large tilt in the usual direction in the southern area. For 1925, the isobases run about east-west; in 1930 the center of uplift was at the outlet of Lake Ontario. In the latest period, about 1935, the tilt was somewhat larger than usual.

In addition to the results of the Great Lakes observations on uplift in the interior of the continent, there are data from the tide gauges at the coast. The material was used in the same way as for Fennoscandia, so that all general remarks concerning Table 4 apply to Table 8. New York with a complete record since 1893 was chosen as the basic station for the Atlantic coast, and San Francisco with data since 1898 for the Pacific coast; the comparison of the change at San Francisco with that at New York is purely formal, as the basic reason for such comparisons, to reduce the effect of meteorological conditions, does not hold in this case. As in Table 4, 10 cm uplift have been added in column 6 to compensate for the eustatic change.

The following general remarks apply to Table 8: At Galveston, the rather rapid sinking of the land has continued (*see* Gutenberg, 1933, p. 461); the previous value of ½ cm per year is reduced to 0.4 cm as a consequence of the correction for the eustatic change. Except for Fernandina, where only a relatively short series of data with rather large changes from year to year is available, the Atlantic coast south of Boston seems to be sinking at present. The corresponding values for the Pacific coast south of Oregon are negative, too, but their absolute values are too small for definite conclusions. Data concerning the tide gauges along coasts of the United States may be found, in addition to the references given previously, with some discussion in a paper by Bowie (1936, p. 23-25).

Unfortunately, the yearly means for most of the Canadian stations do not refer to the calendar year and are sometimes based on partially incomplete series of observations with many years missing. However, for most of them the time interval between the first and the last group is large enough to furnish a rate of uplift with an error probably not exceeding 10 cm per century.

The data for Quebec are of special importance since they improve the connection between the readings of the tide gauges on the Great Lakes and those of points connected with the ocean. Detailed information about the gauge at Quebec has been given by Dawson (1917, p. 13-16). He points out that, in midwinter months and also when the ice moves out in the spring, irregularities may occur. Like the effects of meteorological conditions, these effects probably average out by taking differences between series of years. The following data indicate the accuracy of the resulting rate of uplift at Quebec, using arbitrary zero points:

	1894/1904	1905/1915	1910/1920	1932
		(Feet)		
Sea level at Quebec	0.60	0.51	0.46	0.46
Difference Quebec-New York (F. H.)	0.55	0.48	0.34	0.16

These and other data leave no doubt that the region of Quebec has been rising at a rate of between 30 and 50 cm per century (including again the 10 feet generally added). About the same rate has been reported by A. K. Snelgrove, according to Flint (1940, p. 1776, footnote) for Notre Dame Bay, Newfoundland (50°N., 55° W.).

On the Pacific coast in Canada, only the data for Vancouver and Victoria are sufficient to yield trustworthy results. A slight uplift is suggested for this whole coast as well as for the Alaskan coast. For nine tide-gauge stations in Alaska, Woodworth (1927, p. 73) has compared the sea level about 1890 and about 1920. His results indicate an uplift at all nine points, ranging from about 15 cm per century at Met-

lakatla near the southern boundary of Alaska to more than 2 meters per century at three stations in the Lynn canal. However, there is little doubt that most of these points underwent vertical movements during earthquakes, especially the Yakutat Bay shock in 1899.

Figure 4 shows the uplift in northeastern North America. The isobases are based on the final findings for stations given in Tables 6 and 7. In addition the figure shows the Algonquin hinge line—that is, the zero isobase during the past 10,000 or 20,000 years; the ice front at the end of the life of Lake Algonquin; and the isobases connecting the points showing an uplift of 100 meters and 250 meters, respectively, since the last ice cap began to melt. These four curves have been drawn following Daly (1934). The continuation of the present hinge line to the west is uncertain. It seems to reach the Pacific coast in Oregon or northern California.

All the curves have their center in the region of Hudson Bay. It is very fortunate that since 1928 a tide gauge has been operated at Churchill, Manitoba (Fig. 4), giving data furnished by Mr. F. H. Peters, Chief, Hydrographic Service of the Department of Mines and Resources, Ottawa, Canada. The tide gauge, with a 30-foot-deep channel leading to it, is located on a wharf within the harbor, about 1 mile from the entrance. The average depth across the harbor at this point is 11 feet at lowest tide. The data are affected by a river entering Churchill Harbour.

All monthly mean data available are given in Table 9. The time intervals do not cover calendar months but start with the earliest observation of the year and cover periods of 30 days afterwards. The observations are interrupted during winter by the ice; they begin on April 28 in 1934, on May 1 in 1939, on May 2 in 1937, on May 30 in 1933, on June 1 in 1935 and 1936, and on June 9 in 1938.

As a relatively large yearly period in the mean sea level at Churchill is to be expected because of the effect of the melting of the ice and the flow of the river entering the harbor, the mean change in sea level has been calculated separately for each month. All findings indicate an uplift at a rate larger than 1 meter per century, while for all months except for September the value exceeds 2 meters per century. Therefore we may safely conclude that the present rate of uplift of the coast near Churchill exceeds 1 meter per century and probably is nearer 2 meters per century; even a value of 3 meters per century must be considered as possible since the average rate for the months June to November is close to this rate. There is no reason why the ice or the river flow should produce a spurious increase in the rate of change in sea level in any months, although the data for August and September are probably least affected by these meteorological elements.

It is of interest to compare these results with other findings. A summary of observations was given by Bell (1896). He pointed to many instances, where in course of time an ever increasing difficulty was encountered in reaching establishments from the sea. Driftwood has been

TABLE 9.—*Mean sea level at Churchill, Hudson Bay*

Results are in 1/100 foot above a point 8.62 feet above the Hydrographic Survey Datum. For the time intervals see text. Last two lines: Average fall in sea level, *a* in feet per year, *b* in meters per century.

Year	June	July	August	September	October	November
1928	.....	.....	.....	−11	+23	.....
1929	.....	.....	− 1	+11	+13	+ 1
1930	.....	+ 2	+ 5	+31	+56	+25
1931	.....	+40	+13	+42	+62	+63
1932	.....	.....	+32	+30	+54	+36
1933	+26	−10	+ 6	+23	+43	.....
1934	− 2	− 7	+26	−18	−13	+33
1935	+10	−25	−29	− 2	.....	.....
1936	−42	−40	−26	+ 4	.....	.....
1937	−38	−33	−47	−31	.....	.....
1938	.....	−60	−46	−33	−47	.....
1939	−49	−43	−56	−32	−64	−113
a	0.129	0.085	0.072	0.040	0.099	0.148
b	3.9	2.6	2.2	1.2	3.0	4.5

found along hundreds of miles of coast far above the reach of the highest tides; remains of rotten wood may be detected "up to nearly fifty feet". Stanley (1939) observed it even as high as 85 feet and pointed out that this "suggests the recency of the later uplift and probably also the inactivity of decay in the cold climate." Bell (1896) considered it impossible that the wood had been carried to its present position by abnormally high tides, although these may reach 20 feet or even slightly more during strong gales. He pointed, besides, to conclusions from beach dwellings of Eskimos; on Outer Digges Island (northeast entrance to Hudson Bay), he found old fish traps at an estimated height of 70 feet. He refers also to channels through which ships could pass 200 or 300 years ago, but today have scarcely enough water for a canoe. Little Whale River (east coast near Richmond Gulf) "had to be abandoned as a harbor on account of the increasing shallowness of the water." While nobody doubts the fact that the region of Hudson Bay has been rising in postglacial time, Bell concluded that all facts indicate that the

region is still rising "with no evidence of a contrary character." In earlier papers, Bell had concluded that the waters of Hudson Bay are receding between 5 and 10 feet (1½ to 3 m) in a century and that the rate of uplift near Churchill is about 7 feet (210 cm) per century. This result is confirmed by the very recent tide-gauge data already discussed.

Tyrrell (1896) doubted the findings of Bell. Some of his objections seem indeed to point to a smaller uplift; others are not well taken, as for example, mention of a bridge at a place where now there is no water and "since has been taken by a sandy bar built along the bank of the river by the currents caused by the ebb and flow of the tide." His crucial argument, however, refers to a little bay near Churchill which had been the winter harbor of small sloops. In 1741, two ships entered this bay, one of which seems to have drawn at least 6 feet of water to enter; apparently the maximum is now about 4 feet, indicating a minimum 2-foot rise of the land. On bluffs bordering the bay, names of visitors are cut. Among them are two with the date of May 27, 1753. Tyrrell assumed that on that day the bay was still covered by ice, piled up to the top of the highest spring tide of the winter—about 15½ feet above mean sea level. As these names and others without dates are now about 7 feet above this supposed ice surface, he concludes that an uplift by 7 feet since that time is impossible, that probably the land there "has now reached a stable, or almost stable, condition," and that all other evidence mentioned before "is delusion."

Johnston (1939) has reinvestigated the problem of the bay. In 1934, the names were at an elevation of about 23 feet above sea level. Assuming again that they were written by persons on the surface of the ice, he concluded that little or no uplift has occurred since. Consequently, he argues, the removal of the weight of the ice after the glaciation no longer has any effect in the region of Churchill and, therefore, cannot be the cause of the uplift in the Great Lakes region either. "It seems evident that the recent uplift of the Great Lakes basins is due to some recent and local cause" (Johnston, 1939, p. 98). However, all his conclusions are based on the assumption that on May 27, 1753, the surface of the ice in the bay was 15 or 16 feet above mean sea level. If in 1753 the ice broke up before May 27, and this has happened not infrequently in recent years, the same evidence would lead to an uplift for the past 180 years which might have exceeded 10 feet. Leaving out this uncertain evidence, all information indicates an uplift of the Hudson Bay region noticeably more rapid than in the Lake Superior district.

The center of the uplift and its maximum amount have not yet been determined. Daly (1934, p. 106) considers it likely that the uplift in

the southern part of Hudson Bay, especially about James Bay, has exceeded 300 meters. Stanley (1939) has pointed out that along the east coast "south of Lat. 55°50′ North neither the islands nor the highest hills of the coastal zone are high enough to have recorded the marine limit, and abandoned beaches may be found up to their summits." From spirit levels which were run from tidewater to the "marine limit" at two localities in Richmond Gulf, 18 miles apart in an east-west line, he concluded it likely that "the component of the tilt rate along this line shows a rise of about 3.3 feet per mile westward toward Hudson Bay." [2]

It is also possible to obtain some information as to the change of the rate of uplift with time. Taylor (1926, p. 148) has found an average tilt in the Great Lakes region at the rate of 0.43 feet per mile during the post-Nipissing stage of the lakes. According to Antevs (1939) the corresponding time interval is about 4000 years. This would give an average uplift at the rate of about 2 mm per kilometer per century, as compared with the present rate of about 1 mm per km per century (from Fig. 4). In other words, the present rate of uplift is about half the average rate during the past 4000 years. This result is not highly accurate, as the time interval is not exactly known, and the changes in the lakes probably have affected the uplift; however, it indicates definitely that the rate has slowed down in course of time in a way similar to that in Fennoscandia.

POSTGLACIAL UPLIFT, SUMMARY

Jamieson (1865; 1882) was the first to suggest that the uplift in the previously glaciated areas is due to the tendency toward restoration of equilibrium. Down to the present time doubts have been expressed concerning the correctness of this hypothesis. (*See*, for example, Daly, 1940, p. 316.) The fact that the history and the present details of the uplift in Fennoscandia and in North America are so similar points to the same cause. In both regions the hinge line has remained in approximately the same location for thousands of years and approximately parallels the limit of glaciation. The isobases surround the area of maximum thickness of ice. The speed of the uplift is of the same order of magnitude in both regions and has decreased to about one half during the last 5000 years; the present tilt in the marginal regions of uplift is of the order of 1 mm per kilometer per century. The gravity anomalies are negative with increasing absolute amounts toward the center of glaciation as summarized by Daly (1940, p. 324, 329), indicating that the equilibrium

[2] In his book *Kabloona* (New York, 1941) G. de Poncins states on p. 29 that there was a time when the natives of King William Land (about 1000 km north of Churchill) hunted the whale. He continues: "The land here has risen, the waters have become shallow, the whale comes no more."

has not yet been reached. Finally, such uplift is not restricted to the two regions which have been discussed thus far, but information, although incomplete, is available for other regions which were glaciated during the ice age. The following data are taken from compilations by Daly (1934; 1940), Born (1923), and Woldstedt (1929). In considering them one must keep in mind that the melting of the ice has produced eustatic changes and that local vertical movements are superimposed on the postglacial uplift; both processes had to be eliminated and have introduced inaccuracies in the results.

Simultaneously with the glaciation in Fennoscandia, the British Isles were covered by ice with a center in the Hebrides where the postglacial uplift probably exceeded 100 feet (30 m), decreasing toward Scotland; the zero isobase for the recent milennia intersects northern Ireland and northern England. There is some indication that the zero isobase at present lies south and west of Great Britain (Table 4; Fig. 3). In Spitsbergen the postglacial uplift is about 50 meters or more. In Greenland it increases from zero in the south to hundreds of meters in the north. In Novaya Zemlya about 100 meters have been found. In the Taimyr, Siberia, region the coast seems to be rising at present with indication of uplift during its postglacial history.

In the Southern Hemisphere there are indications of postglacial uplift in South Africa and in South America; the maximum rise found in Patagonia has been about 40 meters. In Antarctica indications of a rise up to 100 meters have been reported as a consequence of the decrease of the ice cap there.

Thus we find the postglacial uplift in practically all regions where it is to be expected. The process is not finished yet. The uplift needed to restore equilibrium in Fennoscandia is about 200 meters in the central zone, as compared with a maximum uplift of about 500 meters in the past. In North America, the figures are probably of the same order of magnitude.

VISCOSITY AND STRENGTH OF THE EARTH

The effect of the forming of the ice during the ice age and later of the melting of the ice on the earth's crust is controlled to a first approximation by three quantities: the compressibility of the layers involved, their strength, and their plasticity. The change in pressure produces a change in volume; theoretically this compression or expansion occurs simultaneously with the cause. An additional weight upon the surface must produce a sinking of the region as a result of the elastic compression of the material below. As the deeper layers of the earth are much less com-

pressible than the surface layers, the effect will decrease with depth. There is no doubt that this purely elastic change amounts to a noticeable fraction of the total change. Daly (1934, p. 129-135) has discussed this problem. He estimated that the amount of the elastic change in level was about 100 meters for Fennoscandia and 160 meters for North America. The effect of a change in weight of a mass having a sinusoidal form has been given by G. H. Darwin. The uplift or depression in the center is given by $ghr\varrho/2\pi\mu$ (g = gravity, h = height of mass in center, ϱ = its density, r = its radius, μ = rigidity in the earth assumed to be constant). If we use for Scandinavia $h = 3000$ meters $= 3 \times 10^5$ cm, $r = 1200$ km $= 1.2 \times 10^8$ cm and an average rigidity of 10^{12} dynes per square cm of the affected layers of the earth, we find an elastic depression by the ice (and after the melting of the ice an elastic uplift) of about 50 meters. This value is to be slightly increased on account of the fact that the rigidity is smaller than 10^{11} dynes per square cm in the upper 50 kilometers, and to be decreased considering the greater rigidity below a depth of 100 km. As the effect of the uppermost layers is probably larger, we reach the conclusion that in Scandinavia the elastic change is probably slightly larger than 50 meters. For North America, we find about 100 meters considering the larger radius of the glaciated area. The true height of the ice may have been noticeably different from the adopted value, and the difference could change the result by 50 per cent.

Fortunately, we can disregard the elastic part of the uplift, if we restrict ourselves to the time after the ice had disappeared. The movement due to "elastic after-working" during the later period of the uplift was relatively small and may be included in the plastic uplift.

The theory of plastic processes is still not well understood. Experiments show that certain materials under given conditions as to temperature and confining pressure do not undergo plastic flow until a certain minimum distortional stress is applied; the quantity called "strength" of the material is equal to this minimum stress. Under stresses greater than the strength the material undergoes plastic changes. On the other hand, "creep" may occur with stresses smaller than the strength. The observed "plastic" flow depends not only on the stress, the temperature, and the confining pressure, but also on the duration of the experiment and on the history of the material; in experiments, irregular behavior is quite commonly found, as a result of the combination of viscous flow, creep, and other physical conditions. For details the reader is referred to Gutenberg (1939) or Griggs (1939; 1940). Goranson (1940) has suggested the use of the expression "mobility" instead of viscosity where plastic flow of solids is involved.

In our problem of postglacial uplift where the data on the uplift as well as the conditions in the interior of the earth are known only to a first approximation, we need not consider complicated physical theories but may make the usual assumption that the strain S in a solid undergoing plastic flow is given by:

$$S = \frac{F}{\mu} + \frac{1}{\nu}\int F dt \quad \ldots\ldots (1)$$

where

$$\nu = \mu\tau. \quad \ldots\ldots (2)$$

μ is the coefficient of rigidity, F the tangential stress, t the time, ν the coefficient of viscosity, and τ the time in which the stress decreases to $1/e$ by plastic flow, if the distortion is kept constant (time of relaxation). ν is measured in "poises"; 1 poise = 1 gram per centimeter per second.

From experiments and from the fact that the mountains have not changed much in historical time, it follows that the outer layers of the earth have a strength great enough to prevent the mountains from flowing apart; for most types of rock under the conditions at the surface of the earth the strength is about 10^9 dynes per square centimeter or slightly more. On the other hand, the fact that we observe the postglacial uplift under stresses (indicated by the gravity anomalies) considerably smaller than those produced by the mountains leaves no doubt that the strength in deeper layers of the earth must be very small. The fact of isostasy (excluding regions where tectonic processes are acting) leads to the same conclusion and, according to Heiskanen (1940, p. 168) and Tsuboi (1940, p. 453), points to a depth of about 60 km for the transition from the layers with large strength (lithosphere) to the deeper asthenosphere in which the strength should be much smaller to permit a "floating" of the upper layers. Unfortunately, the assumption of a homogeneous, plastic earth without strength leads to such complicated equations that no attempt has yet been made to introduce an additional upper layer with great strength which could be assumed to be infinite.

The simplified problem of viscous flow in a homogeneous incompressible earth without strength has been attacked in various ways. Haskell (1935) gave a fundamental formal solution for the motion of a highly viscous fluid when a symmetrical pressure is applied to the surface and applied the solutions to the postglacial uplift. In a second paper (1936) he treated the case of a load in the form of an infinitely long strip with parallel sides, and in a later paper (1937) he summarized and extended his previous results. At the time t_o (removal of the ice load) the rate of uplift $v_{r,o}$ at a distance r from the center of glaciation is purported to be given by:

$$v_{ro} = v_o e^{-b^2 r^2} \quad \ldots\ldots (3)$$

where v_o and b are constants to be determined from the rate of uplift at the time t_o. If d_o is the displacement (downwarp) at the center for $t = 0$, g = gravity, ν = viscosity as defined by equation (1), he finds:

$$\nu = \frac{g\rho}{2b\sqrt{\pi}} \frac{d_o}{v_o} = C\frac{d_o}{v_o}, \qquad \text{where } C = \frac{g\rho}{2b\sqrt{\pi}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (4)$$

All quantities are to be measured in cm, gr., sec. (1 year = 3.16×10^7 seconds). In applying this equation to the conditions in Fennoscandia,

TABLE 10.—*Data concerning the uplift in Fennoscandia*

Year	v (cm/year)	b (per km)	D (meters)	ν (poises)
6800 B. C.	13	.....	250	0.5×10^{22}
6000 B. C.	6.7	.....	180	0.7
5000 B. C.	3.9	0.00127	147	1.2
4000 B. C.	2.7	0.00119	118	1.7
3000 B. C.	2.2	0.00124	94	2.0
2000 B. C.	1.8	0.00125	74	2.4
1900 A. D.	1.1	0.00125	0	.....

he used data given by Nansen concerning the uplift. According to a written communication, he added 20 meters for the future uplift, following a suggestion made by Nansen. As the true amount of the uplift still to be accomplished seems to be about 200 meters, his values of d are too small by about 180 meters. This was pointed out by Meinesz (1937) who added 180 meters to the uplift given by Haskell and found from Haskell's data ν = about 3×10^{22} poises.

Apparently it has not been realized that it is possible to calculate the order of magnitude of the viscosity without an assumption regarding the future uplift d_o. We suppose with Haskell that equation (3), which is correct only for the time $t = 0$, is also a good approximation for later times. Haskell found that under his suppositions the value of b is practically constant if it is calculated from equation (3) for various times (Table 10). At the present time we have the deviation $d = d_2$ and the rate of uplift v_2. At an earlier time the corresponding values are $d_1 = D + d_2$ and v_1, where D is the uplift from the earlier time until the present. Applying equation (4) to both times, we have:

$$\frac{d_2}{v_2} = \frac{D + d_1}{v_1} = \frac{D}{v_1 - v_2} \text{ and } \nu = \frac{d_2}{v_2} C = \frac{DC}{v_1 - v_2} = \frac{2.3\, D \times 10^{20}}{v_1 - v_2}$$

if we assume that the density $\rho = 3.3$, and that $b = 1.25 \times 10^{-8}$ per cm (Table 10); D is to be measured in meters and v in cm per year. The

results are given in the last column of Table 10, of which the first two lines are added from Figure 5, and which otherwise is based on the data of Haskell (1937, p. 24).

The results show an apparent increase of the coefficient of viscosity in course of time. In interpreting this result we must remember that supposedly the viscosity does not change with depth or time and that the velocity of flow is proportional to the shearing stress. The result that the calculated viscosity increases with time may be a consequence of the fact that in Haskell's theory equations (3) and (4) are correct only for $t = 0$, but have been used by us for later times. As we shall see, Niskanen (1939) does not follow this procedure and arrives at results which do not depend on the time of the observations. However, it may be that in addition there is a real change in the coefficient of viscosity (as defined). Griggs (1939; 1940) has found from laboratory tests that creep or pseudoviscosity differs from ordinary viscosity in that the velocity or flow increases about exponentially with the stress. Thus, the calculated increase of ν may be a consequence of the fact that the speed of the plastic flow decreases faster than proportionality with the stress. Of course, equations (1) and (2) would not be correct in this case. We must, finally, remember that the upper layers, contrary to our assumptions, have a relatively great strength. We cannot decide how much each of these factors contributes to the result. However, there can be no doubt that the modified equations of Haskell furnish a good first approximation.

The order of magnitude found for the viscosity corresponds approximately with the values found in laboratory experiments. (*See*, for example, Griggs, 1939, p. 235; 1940.)

The coefficient of rigidity in the outer part of the mantle of earth is about 10^{12} dynes per cm^2. Using this value and the viscosity $\nu = 3 \times 10^{22}$ poises, for the present time, we find from equation (2) that the time of relaxation τ is about 3×10^{10} seconds or about 1000 years. This figure gives the time in which the stress decreases to $1/e$ by plastic flow if the distortion is kept constant, and should not be confused with the time ϑ in which the distortion is reduced to $1/e$ if the load is removed. The relationship of these two constants depends on the conditions; however, in general their order of magnitude does not differ much. τ is a constant of the material; ϑ, in addition, depends on the process and may change in time.

One of the most important conclusions from Haskell's investigations is that the plastic flow extends to relatively great depths, as has been suggested by Daly (1934). According to Haskell's findings, at a depth equal to one half the width of the area of uplift (about 1000 km in Fenno-

scandia) the maximum vertical displacement has 0.8 of its value at the surface and at twice this depth remains more than one half, assuming that the viscosity does not change with depth. The result indicates that the plastic flow which tends to compensate the movements near the surface in the regions of postglacial uplift extends downward to the surface of the core and probably even affects the core.

Van Bemmelen and Berlage (1935) in a more general investigation on plastic flow in the earth have treated the problem of the postglacial uplift differently. They suppose that the sinking due to the weight of the ice had occurred along a trough. In Fennoscandia, the axis of this trough would have to be assumed in a southwest-northeast direction. The original depression along the axis is d_o and is supposed to decrease with the distance x from the axis following the form $d_o e^{-b^2x^2}$, which is similar to the one assumed by Haskell. The fundamental assumption of van Bemmelen and Berlage is that the movements can be treated as a part of a wave of a highly viscous fluid which has a depth h. Below this depth all movements are supposed to be negligible. The depression of the points of the axis at the time t (since the beginning of the uplift) below the position of equilibrium is proportional to $e^{-c\rho t}$, where:

$$c = gh^3/3\nu \qquad (5)$$

ρ = density, ν = viscosity, h = depth to which viscous material extends, g = gravity. The following values are used in evaluating the equations (for details *see* van Bemmelen and Berlage, 1935): uplift in center to date, 280 m; original rate of uplift, 14 cm per year; present rate, 1.1 cm per year. The following values result: original depression in center, 490 meters (or uplift of 210 meters still to be expected); time elapsed since beginning of uplift, 9500 years; $b = 0.0017$ per km, and finally $c = 4.26 \times 10^3$ (cm, gr., sec.). The last step is now to use equation (5) to find the viscosity ν. Van Bemmelen and Berlage believed that all plastic movements involved in the process are negligible at a depth greater than 100 km, and, assuming $h = 10^7$ cm find $\nu = 1.3 \times 10^{20}$ poises. However, the viscosity seems to be better known than the depth to which the plastic deformation extends with an appreciable amplitude, and therefore the author considers it preferable to assume an average viscosity and calculate h from equation (5). Taking $\nu = 3 \times 10^{22}$ poises, we find $h =$ ca 700 km. This result means that the processes connected with the uplift are acting with relatively large amplitudes to a depth of many hundred kilometers. As the method is based on hydrodynamic equations with assumptions completely different from those used by Haskell, and as an error by a factor 10 in assuming the viscosity changes the resulting

h by little more than a factor 2, the conclusion seems inevitable that the movements extend to many hundreds of kilometers downward with scarcely decreasing amplitudes.

Once again Meinesz (1937) offered a very useful first approximation to the problem. He estimated the negative anomaly in the central part of the depressed area to be about —25 milligal, and dividing this value by $2\pi\kappa\rho$ (κ = Newton's constant = 6.67×10^{-8} cgs) he found 180 m for the remaining uplift. This result is somewhat uncertain on account of possible errors in the assumed mean anomaly. It agrees well with the findings of the preceding and the following methods.

Meinesz assumed that the rate of uplift (v) is always proportional to the deviation d from the equilibrium surface, or that:

$$d = d_0 e^{-Kt} \quad \text{or} \quad K = -\frac{v}{d} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (6)$$

Meinesz himself has pointed out that this is only an approximation. He used data by Nansen (1928) as given by Haskell and added 180 meters to find d. However, since Haskell's values are 20 meters greater than those given by Nansen, he really added 200 meters. Using equation (6) and combining the data with his value of $d = 180$ meters for today, he finds $K = 2.76 \times 10^{-12}$ per second. Since this makes the observed rates for the early times too small and for the second half of the period too large, he finally adopted the value $K = 2.3 \times 10^{-12}$ per second or 7.3×10^{-5} per year. Of course, this does not remove the discrepancies. Meinesz considers that the horizontal dimensions of the area of uplift may have decreased with the passage of time (which is not supported by the observations) or that other processes may be involved. As we shall see, the increase in the value of K corresponds with the theory of Niskanen. From equation (6) it follows that for two times t_1 and t_2, with the deviations d_1 and d_2 from the equilibrium, the quantity K (which is supposed to be constant with time) is given by:

$$K = (\log_e d_1 - \log_e d_2)/(t_1 - t_2). \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (7)$$

On the other hand, K can be calculated from equation (6). The essential data are given in Table 11. The values of d are taken from Nansen, and 200 meters is added for the future uplift. The values for d are the same as those used by Meinesz except for the last. They are plotted in Figure 5, and the rate of uplift v was determined from the tangents to the curve (Table 9; Fig. 5). K was calculated in three different ways, (a) from equation (7) using succeeding values of d; (b) by using the value of d for the given year and the last value $d = 200$ meters for 2000; (c) from equation (6) $K = -v/d$. All indicate clearly the decrease of K with time contrary to the assumption v/d = constant. Finally the rate

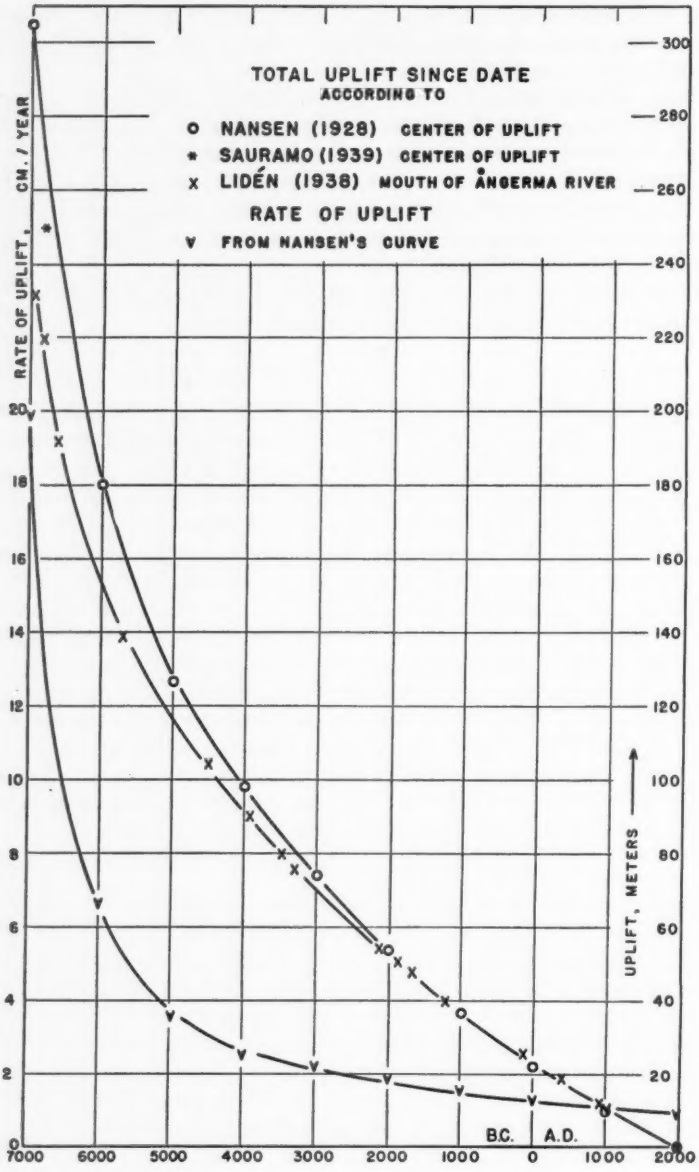


FIGURE 5.—*Uplift and its rate in Fennoscandia as a function of time*

of the uplift *(v)* was calculated from equation (6) using the value $K = 0.73 \times 10^{-4}$ per year found by Meinesz. This corresponds approximately to the value of K for the year 2000 B. C. For earlier years, a larger value has to be taken to find the observed v, and for later years

TABLE 11.—*Remaining uplift d, rate of uplift v, and constant K*

Based on data given by Nansen (1938) for the central part of the region of postglacial uplift in Fennoscandia.

Year	$-d$ (meters)	v (cm/year)	$\log_e d$	K per 10,000 years			v calc. (cm/year)	$\frac{-d}{v^{1/3} \times 10^4}$
				equation (7)		$-v/d$		
				a	b	c		
7000 B. C.	505	(20)	6.225			4.0	3.7	1.9
				2.85	1.03			
6000 B. C.	380	6.7	5.940			1.8	2.8	2.0
				1.50	0.80			
5000 B. C.	327	3.8	5.790			1.2	2.4	2.1
				0.93	0.70			
4000 B. C.	298	2.6	5.697			0.87	2.2	2.1
				0.84	0.67			
3000 B. C.	274	2.2	5.613			0.80	2.0	2.1
				0.76	0.63			
2000 B. C.	254	1.8	5.537			0.71	1.8	2.1
				0.69	0.60			
1000 B. C.	237	1.5	5.468			0.63	1.7	2.1
				0.65	0.57			
0	222	1.3	5.403			0.59	1.6	2.0
				0.56	0.52			
1000 A. D.	210	1.1	5.347			0.52	1.5	2.0
				0.49	0.49			
2000 A. D.	200	1.0	5.298			0.50	1.45	2.0

a smaller value. Again, we see that the uplift is retarded, and the discussion of this fact in connection with the findings based on Haskell's theory applies here also. The last column of Table 11 shows that not d/v, but $d/v^{\frac{1}{3}}$ is approximately constant. This agrees with Niskanen's equations to be discussed later.

Supposing that $K = 0.73 \times 10^{-4}$ per year, the time T in which half of the anomaly disappears is given by $T = 0.693 \,/\, K = 9500$ years; again this value corresponds approximately to the time 2000 years B. C. According to our results, the uplift was more rapid before that time and is slower now; in the future it will take more than 10,000 years to reduce

the remaining deviation from the equilibrium to one half or to accomplish about 100 meters of additional uplift in the center of the region.

For his further calculations Meinesz assumed that the uplift occurs along a trough and that the displacements are part of a sinusoidal curve perpendicular to the direction of the axis. If $2R$ is the diameter of the trough, ϱ the density of the material, v again the rate of uplift and d the vertical displacement, he finds:

$$v = \frac{\rho g R d}{\pi \nu} = -Kd \quad \text{or} \quad K = \frac{\rho g R}{\pi \nu} \qquad (8)$$

With the values of Meinesz, $K = 2.3 \times 10^{-12}$ per second, $r = 700$ km, and a density of 3.3 for the substratum, equation (8) gives $\nu = 3 \times 10^{22}$ poises.

Meinesz finally tried to calculate the change of viscosity with depth. Assuming an exponential law for the viscosity ν as a function of depth and proceeding otherwise as above, he finds a decrease of ν from 3×10^{23} poises at a depth of 150 km to 3×10^{22} near the boundary of the core of the earth. However, we must consider that the accumulation of assumptions and especially the possibility that the viscosity inside the core may be very small casts some doubt on the result.

Previously we have pointed out that the time of relaxation of the material (τ) must not be confused with the time $\vartheta = \frac{1}{K}$ in which the displacement in the center of the region decreases to $1/e$. A comparison of equations (2) and (8) shows that Meinesz' assumptions leads to $\tau/\vartheta = \varrho g R/\pi\mu$ and that the ratio of τ and ϑ depends only on the radius R of the area of uplift for a given material. Using $R = 1000$ km and supposing an average rigidity of 10^{12} dynes per cm^2, we find the following orders of magnitude for Fennoscandia: τ = about $10^{-9} R\vartheta$ = about $0.1.\vartheta$ As ϑ = about 14,000 years, τ is the order of 1000 years. The time in which the deviation from the equilibrium reaches $1/e$ (or any other fraction, such as one half) of its original value is inversely proportional to the horizontal extent of the glaciated area and is not affected by the height of the original load. This result depends, of course, on Meinesz' assumptions as to the process involved.

Finally, Niskanen (1939) has discussed the facts concerning the uplift in Fennoscandia as well as the theory. He makes assumptions similar to those of Haskell and applies equations for a viscous fluid. The deviation d from the equilibrium is given by:

$$d = \frac{d_o}{\sqrt{1 + 4kt}} e^{-\frac{b^2 r^2}{\sqrt{1 + 4kt}}} \qquad (9)$$

where d_o is the deviation in the center at the time $t = 0$, r = distance from the center (axis) of glaciation, b is the same constant used previously which controls the change in the direction perpendicular to the axis, and k is a new constant related to the speed of the uplift. The rate of uplift in the center is given by

$$v = 2\,k\,d_o\,(1 + 4\,kt)^{-3/2} = v_o(1 + 4kt)^{-3/2} \quad \ldots\ldots\ldots\ldots\ldots\ldots (10)$$

At the time $t = 0$, there is $v_o = 2kd_o$.

Whereas in the preceding solutions except for the general solution given by Haskell the time entered only as an exponential function, so that any year could be taken as zero for the time, this is no longer true for the solution given by Niskanen. In the center of the uplift the ratio of the deviations d_1 and d_2 at the times t_1 and t_2 is given by $d_1^2/d_2^2 = (1 + 4kt_2)/(1 + 4kt_1)$ as may be found from equation (9) for $r = 0$. Contrary to the preceding solutions the time T, in which the remaining uplift decreases to half the amount at a given time, is no longer constant but $T = (3/4\ k + 3t$ increases with the time t since the beginning of the process. This fact corresponds to the observations, as we have seen. Only, at the moment when the ice disappeared (supposing that the melting occurred fast enough not to permit a noticeable plastic flow), does one find the equation given by Niskanen: $T = 3/4\ k$.

Unfortunately, the time before 6800 B. C. cannot be used for calculations, as Niskanen himself has pointed out. Before 6800 B. C., the elastic uplift and tilt due to other processes, for example the changing level of the Ancyclus Lake, had been superimposed on the plastic uplift. Moreover, the theory supposes a very rapid (sudden) melting of the ice, so that the rate v_o at the beginning of the uplift cannot be found with any great accuracy.

In order to find k from equation (10), Niskanen used the data given by Lidén (1938) on the uplift in Angermanland (Fig. 5). For the year 6800 B. C., Niskanen assumes a rate of uplift of 13 cm per year, and for the present time a rate of 1.2 cm per year; he finds $k = 1.07 \times 10^{-4}$ per year, T = about 7000 years. Apparently he did not attempt to correct for the fact that the year 6800 B. C. was clearly not the time $t = 0$ to be used in his equations.

The main advantage of Niskanen's method is that it permits the calculation of the remaining deviation from the equilibrium and the total depression of the land at the time when the ice began to melt. For the center of the area ($r = 0$) the exponential term in equation (9) equals 1, and the uplift $D = d_o - d$ since the time $t = 0$ is given by:

$$D = d(\sqrt{1 + 4kt} - 1) \quad \ldots\ldots\ldots\ldots\ldots\ldots (11)$$

from which the future uplift d may be calculated. Niskanen assumes an uplift of $D = 250$ meters in the central area since the disappearance of the ice; this amount of uplift has been found by Sauramo (1939) since the year 6800 B. C. His results are $d = 210$ meters for the future uplift and a depression of the central area by 460 meters at the year 6800 B. C.

The results must be considered only as an approximation, since Niskanen has assumed that the year 6800 B. C. is the zero point of his time scale. Fortunately, it is possible to find better approximations if we introduce the zero point of his time scale as an additional unknown quantity. From equation (10) it follows that $(1 + 4kt)^{-\frac{1}{2}} = (v/v_o)^{\frac{1}{3}}$. If we introduce this quantity in equation (9), leaving out the exponential term for the center, we find $d/v^{\frac{1}{3}} = d_o/v_o^{\frac{1}{3}} =$ constant, as d_o is the total depression at the center of glaciation at the beginning of the uplift, and v_o the rate of uplift at this time; of course, v_o is only of theoretical interest. Contrary to the assumptions, the uplift had begun gradually. The quantity $d/v^{\frac{1}{3}}$ has been calculated from the observations (Fig. 5) and added in the last column of Table 11; it is practically constant and shows that the equations of Niskanen are an excellent approximation. If the rate of uplift is measured in cm per year, and the remaining uplift d in meters, we have $d =$ about $200\ v^{\frac{1}{3}}$. The deviation d_1 at a time t_1 (rate of uplift v_1) and the deviation $d_2 = d_1 + a$ at an earlier time t_2 (rate of uplift v_2) are connected by the equation:

$$d_1 = \frac{a v_1^{\frac{1}{3}}}{v_2^{\frac{1}{3}} - v_1^{\frac{1}{3}}}.$$

This equation permits the calculation of the future uplift d_1 at any given moment. The following values for v and a are taken from Figure 5:

Year	v in cm/year	Difference a in uplift down to 1900, in meters	Calculated future uplift d for 1900 in meters
6800 B. C.	13	250	190
5500 B. C.	5	150	210
3500 B. C.	3	85	190
1900 A. D.	1	0	

The average is about 200 meters. With this value, it is possible to find approximately the depression in the central area at the beginning of the uplift: About 250 meters had occurred before 6800 B. C. (data given by Niskanen, 1939, p. 28), 250 meters between 6800 B. C. and the present day, and about 200 meters still remain for future rising, giving a total of 700 meters which agrees well with other estimates. However, not all this uplift is due to plastic uplift; as we have seen, between 50 and 100 meters are due to elastic changes. Consequently, the value d_o in our equations is about 600 meters. This may now be used to calculate the

theoretical rate of uplift at the time $t = 0$. We use the equation $v_o = v\left(\frac{d_o}{d}\right)^3$ where v and d are the corresponding values of the rate of uplift and the uplift still to be expected at a given time, and find:

Year	6800 B. C.	5500 B. C.	3500 B. C.	1900 A. D.
d	450	350	285	200 meters
v	13	5	3	1 cm per year
v_0	30	25	28	27 cm per year

with an average of $v_o = 27$ cm per year. The agreement is better than one would expect. We can also use the fact that the last column of Table 11 gives $d_o/v_o^{\frac{1}{3}} \times 10^4 = 2.0$, or $v_o = 27$ cm per year.

We can now calculate the constant k. $k = v_o/2d_o = 2\frac{1}{4} \times 10^{-4}$ per year. The time t_o can be found from equation (10), using the same values v as above:

Year	6800 B. C.	5500 B. C.	3500 B. C.	1900 A. D.
kt	0.16	0.52	0.83	2.0
t	700	2300	3700	8900 years
t_0	7500	7800	7200	7000 B. C.

Thus the year $t = 0$ of our calculations was about 7500 B. C. As in the previous findings based on Niskanen's method we notice the good agreement between the results regardless of the year which is used, so that we may safely conclude that his equations (regardless of their physical interpretation) are a better approximation than those discussed before and that the time T, in which the deviation from equilibrium is decreased to one half, is increasing with time. Our results indicate $T =$ 4000 years at the beginning of the uplift, $T =$ 16,000 years in the year 3000 B. C., and $T =$ about 30,000 years today.

We also have confirmed the previous findings that the remaining uplift is about 200 m. The gravity anomaly $\triangle g$ which corresponds to this deficiency in mass at present is given by $\triangle g = 2\pi\varrho\kappa d$; $\kappa = 6.67 \times 10^{-8}$ cgs. Consequently $\triangle g =$ about -30 milligals. This value corresponds well with the observations, as may be seen from the map given by Niskanen (1939) or Daly (1940, p. 325). However, we must bear in mind that other effects may produce a large fraction of the anomalies.

Thus far we have dealt with the uplift in the center of the glaciated area. If we investigate the uplift at the distance r from the center, the quantity b enters in the various equations. Meinesz has used a ratio $e^{-b^2r^2}$ for the rate of uplift (velocity) v at the distance r and the velocity v_c at the center for any given time. Haskell has assumed this same ratio for the time $t = 0$ but did not calculate its change with time ex-

plicitly. In the theory of Niskanen this ratio v/v_c decreases with time and is given by:

$$v/v_c = (1 - \frac{2b^2r^2}{1 + 4kt}) e^{-\frac{b^2r^2}{1 + 4kt}}$$

$$\text{For } t = 0 \text{ we have } v/v_c = (1 - 2\, b_2\, r_2)\, e^{-b^2r^2}.$$

Niskanen (1939, p. 21) has used the velocity of uplift at about the year 6800 B. C. (examples in Table 5) to find the quantity b, supposing that the year 6800 B. C. corresponds to the time $t = 0$, and the second of the two equations above can be used. This procedure gives a value of b which is slightly too small. His result $b = 0.00131$ per kilometer agrees well with the similar values (0.0017, 0.0012) of the other authors mentioned which has been found by use of completely different methods and observations. The following data give the order of magnitude of v/v_c using data mentioned before:

	v_c	Velocity v in cm per year			v/v_c		
r =	0	250	400	600 km	250	400	600 km
Year							
6800 B. C.	13	9	3	1½	0.7	0.23	0.12
1900 A. D.	1	0.8	0.5	0.25	0.8	0.5	0.25

The results show a clear increase of v/v_c with time, however, the theoretical change for v/v_c assuming Niskanen's equations are larger than those indicated by the observations; according to Niskanen's results, the rate of uplift should approach the same value over the whole region faster than it really does. On the other hand, the profiles calculated by Niskanen (1939, p. 22) and corrected for the effect of the time before 6800 B. C. agree well with the observations.

If $(1 + 4kt) = 2b^2r^2$, the equations give $v = 0$. Thus we may find the radius of the zero isobase. It increases with the time t. If r is greater than this critical value, v/v_c is negative; however, its absolute value cannot increase much with increasing r due to the exponential term. The details depend on the constants, especially the value of b. Nevertheless, it can be stated that the equations predict that the zone of postglacial uplift should be surrounded by an area of slow sinking of the land and that the line of no change (zero isobase) increases its distance from the center of uplift in course of time. Observations agree with the theoretical results (Figs. 3, 4). The zone of sinking of the land is also found in the theories of Haskell (1936, p. 61), van Bemmelen and Berlage (1935, p. 41), and Meinesz (1937). The details depend on the basic assumptions of these theories as well as on the selected constants. On the other hand, the rate of sinking is so slow that local processes affect the

observations and locally may even produce an uplift where the plastic flow due to the disturbed equilibrium after the ice age would result in a sinking of the land. For these reasons, a detailed comparison between the findings using the various theories and the observations does not promise trustworthy conclusions.

Under the assumptions of Niskanen, the viscosity ν is given by:

$$\nu = \frac{\rho g}{2kb} = \frac{\rho g}{b}\frac{d_e}{v_e} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (12)$$

In comparing this equation with Niskanen's formula it is to be considered that he used the "kinematic viscosity" which is the coefficient of viscosity used here divided by the density ρ. He finds a kinematic viscosity of 1.1×10^{22} poises; this gives $\nu = 3\frac{1}{2} \times 10^{22}$ poises.

If we use our value $k = 2\frac{1}{4} \times 10^{-4}$ per year, or 7×10^{-12} per second, $b = 1.3 \times 10^{-8}$ per cm, and $\rho = 3.3$, we find $\nu =$ about 2×10^{22} poises in good agreement with the values found from the other methods.

At the beginning of the uplift, the ratio of the two quantities τ and ϑ discussed previously is given by $\frac{\tau}{\vartheta} = \frac{4k}{e^2 - 1}\frac{\rho g}{2k\mu b} =$ about $\frac{g}{\mu b} =$ about 0.1. This is exactly the value which we had found using the theory of Meinesz. However, ϑ increases with time, whereas τ is a constant of the material.

If we summarize all the findings on the uplift in Fennoscandia, we must first emphasize that the agreement between the results found in different ways by using different data is good and practically proves that the uplift is isostatic and due to deficiency in mass after the melting of the ice. The results strongly support Daly's hypothesis (1925) that the plastic flow, which furnishes the material for the uplift, extends well below the 1000-kilometer level with amounts not much less than the movements at the surface. The time required to restore half the deviation from equilibrium is of the order of magnitude of 10,000 years and increases with time. If one assumes that the whole mantle of the earth has no strength and everywhere has the same coefficient of viscosity, this is found to be about 2×10^{22} poises. If the effect of the upper 60 kilometers with strength great enough to prevent any plastic flow is considered, it still seems likely that a viscosity of the order of magnitude of 10^{22} poises will be found; moreover, this is the order of magnitude for the viscosity of the few types of rocks which have been investigated in laboratories. The slight decrease of viscosity with depth found by Meinesz must be interpreted with the fact in mind that he has assumed a continuous change with depth; his result may mean that the viscosity decreases gradually to the center of the earth; or the viscosity may increase slightly with depth in the mantle down to the core of the earth

and have a much smaller value inside the core. This latter distribution would correspond to the change of the rigidity with depth.

No calculations have been based thus far on the uplift in North America. This has two reasons: first, the data are more scanty; and second, the process was much more disturbed there than in Scandinavia by relatively large physiographic changes, especially formation of lakes with variable extension and depth. (*See*, for example, Antevs, 1939.) However, all essential quantities differ only by small factors from the corresponding quantities in Fennoscandia, so that we may expect that the observations lead to approximately the same conclusions. For example, if we suppose a maximum rate of uplift of 200 cm per century and a radius of 1500 km of the zero isobase, we find the quantity b of the order of 0.001 per km or 10^{-8} per cm, about the same as in Fennoscandia. If, on the other hand, we make the reasonable assumption that the equation $d = 200v^{\frac{1}{2}}$ (d in meters, v in cm per year) which we have found for Fennoscandia is a good approximation for the conditions in the Hudson Bay region, we find for the remaining uplift $d =$ about 250 meters, only slightly larger than in Fennoscandia. Consequently it is to be expected that the depth and size of Hudson Bay will decrease rather rapidly in the future, and a little of the northern part if anything will be left of the Hudson Bay when the uplift approaches its completion. The decrease in load in Hudson Bay corresponding to the loss of a layer of water in the average at least 1 meter thick during one century must increase the uplift. A similar but smaller effect is to be expected in the Gulf of Bothnia and may be one of the reasons for the fact that the center of uplift there at present is slightly farther to the north than during the beginning of the process. Niskanen (1939, p. 24) has already pointed out that after about 2600 years there will be dry land in the middle part of the Gulf of Bothnia between Finland and Sweden, if the process continues undisturbed; and 10,000 years hence there will be only a small lake where there is now the Gulf.

As the deviation from the equilibrium in the region of postglacial uplift in North America seems to be slightly larger than in Fennoscandia, we should expect a similar ratio for the gravity anomalies. Unfortunately, there are no data for the Hudson Bay region. In the Great Lakes region the average anomalies are about −15 milligals (*see* Daly, 1940, p. 330) as they should be, but the values scatter very much. The following are Airy-Heiskanen anomalies according to Heiskanen (1939) for points north of Lat. 48° between Long. 88° and 72° West. They are based on the international formula. Use of the Heiskanen triaxial formula would increase the numerical values by a few units and give larger anomalies.

The average of −21 milligals, or even about −25 milligals, based on the triaxial formula of Heiskanen (*see* Daly, 1940) would correspond to an uplift of 150 to 200 meters still necessary in this region to restore equilibrium. This seems too large rather than too small for this region which is almost 1000 kilometers from the center of glaciation.

Station	West Long.	North Lat.	Anomaly, milligals
Schreiber	87.3	48.8	−13
Nakina	86.7	50.2	−13
White River	85.3	48.6	−26
Hearst	83.7	49.7	− 4
Cochrane	81.0	49.1	−27
Taschereau	78.7	48.7	−42
Robertval	72.2	48.5	−21

Some information may be found from the following equations. If we suppose equation (6) (Meinesz' assumption) as correct, we may derive:

$$v = -Kd_o e^{-Kt}, \text{ or } \log(-v/d_o) = \log K - 0.434\,Kt. \quad \ldots\ldots\ldots\ldots (13)$$

The minus sign in the parenthesis is due to the fact that the velocity v is directed upward and the displacement d downward. The quantity 0.434 enters because of the transition from natural to Briggs logarithms. K is a small fraction, and therefore log K is negative. If K is very small, the first term on the right-hand side has a numerically large negative value; if K is large, this is true for the second term. The absolute maximum of the right-hand side occurs for $K = 1/t$, and, consequently, there is always $\log(-d/v_o) \geqq \log t + 0.434$. If we apply this equation to the past 10,000 years, $\log(-d_o/v) \geqq 4.43$; $d_o \geqq 27{,}700\ v$; the absolute value of d_o must be at least $277v$, where d_o is the deviation from the equilibrium in meters at the beginning of the period and v the present uplift in cm per year. For Fennoscandia, we have $v = 1.1$ cm per century, and consequently the deviation from the equilibrium 10,000 years ago must have been at least 300 meters. In the Hudson Bay region d_o is possibly at least 500 meters corresponding to $v = 2$ cm per century.

This method supposed that the deviation from the equilibrium in the center decreases exponentially with time. Another restriction is that t cannot exceed the time since the ice had disappeared. If this moment is taken as $t = 0$, the symbol "=" has to be used in the equations.

Equation (13) not only gives us a minimum for the deviation from the equilibrium but also limits for K. If we take $t = 10{,}000$ years, then in Fennoscandia as well as in the Hudson Bay region $\log(-v/d_o)$ lies between −4 and −5, and K must be less than 10^{-3} and over 10^{-5} per year to satisfy the equation. Since the time in which the anomaly is

reduced to one half is given approximately by $T = 0.7/K$, in both regions, the plastic flow must continue for at least about 1000 years, but less than 70,000, to restore half the equilibrium. For Scandinavia we have found the more exact value of about 10,000 years; for the Hudson Bay region, this method at least gives the limits just mentioned.

CONCLUSIONS

The general results, in combination with other data, allow us to get a better idea of the facts which must be considered in interpreting tectonic processes. So far as is known, the visible mountains have not changed their height appreciably in historic times. More data bearing on the accuracy of this statement are very desirable since the experiments of Griggs (1939) on "creep" of rocks suggest the possibility of changes during long periods. The greatest differences in elevation between near-by points of the earth's surface do not exceed about 10 kilometers producing stresses just below the strength as determined by laboratory experiments. Gravity determinations lead to the conclusion that in general the weight of the mountains or the deficiency of material in oceanic regions is compensated within the uppermost layers of the earth, not over 60 km thick, except for certain regions with gravity anomalies. However, in most of these exceptional regions earthquakes occur rather frequently indicating that tectonic processes are going on in such regions, either with a tendency to restore equilibrium or as a consequence of processes which maintain or increase the anomalies. The strength of the material in the uppermost layers down to a depth of about 50 km prevents plastic flow as long as the stresses are less than about 10^{10} dynes per square cm, and must lead to earthquakes as soon as the breaking strength of the material is exceeded by the stresses. Tsuboi (1940, p. 454) has estimated the maximum energy which can be stored in the earth's crust without disturbing isostasy beyond the observed values and found an excellent agreement with the maximum energy liberated in earthquakes (10^{25} ergs).

Whereas at present we have no observations proving plastic or "pseudoplastic" flow in relatively large sections of the lithosphere, the postglacial uplift leaves no doubt that some kind of flow must occur in the deeper parts, say below 60 or 80 kilometers. Seismic observations lead to the conclusions that at a depth of about 70 km there is a small but distinct decrease in the wave velocity, which has been interpreted as a consequence of the transition from the crystalline to the vitreous state. This would explain the fact that at about this depth the strength of the material decreases noticeably and in the asthenosphere below permits plastic flow even if the stresses are small. However, the speed of this flow is very slow. It takes thousands of years to reduce deviations from

the equilibrium to one half. If the stresses accumulate at a rate much faster than they are reduced by plastic flow, the breaking strength of the material will be reached practically as quickly as in the lithosphere, and earthquakes must result in just the same way. Since in certain regions the time interval between earthquakes is measured by centuries or even years, as in the Hindu Kush region at a depth of over 200 km, the possibility of rapid accumulation of stresses due to unknown processes is proved. Thus far there is no contradiction among the observations, the conclusions from laboratory experiments, and the results of the theory based on the assumption of the properties at the various depths which we have just outlined. In particular it is not necessary to look for an explanation for the "deep focus earthquakes" as a spurious phenomenon; all observations on these shocks fit the theoretical conclusions excellently, and there is practically no possibility of explaining the observations on the assumption of a focus at a depth of less than 50 km; on the other hand, contrary to the opinion held by a few, there is no reason why tectonic earthquakes should not occur at greater depths than 50 km. On the contrary, the question must be raised why observations do not indicate their occurrence at depths greater than about 700 kilometers. Do the forces which produce the stresses act mainly in the upper few hundred kilometers? Or does the viscosity at a depth of about 700 kilometers decrease to values below the order of 10^{20} or 10^{19} poises, allowing so rapid a plastic flow that stresses are reduced to one half within a few years or less? Or does the breaking strength increase at a depth of about 700 km so rapidly that below this depth earthquakes are practically prohibited? Findings by Goranson (1940) indicate that beyond a certain confining pressure the breaking strength increases rapidly; the relatively great energy of many very deep shocks would also agree with this possibility.

Finally, it is not unreasonable to assume that inside the core of the earth the viscosity, as well as the rigidity, is much less than in the mantle and permits relatively rapid plastic flow.

More far-reaching conclusions may be drawn. The deficiency in mass in the regions of postglacial uplift corresponding to gravity anomalies of about 30 milligals produces plastic flow throughout the mantle of the earth with only slightly decreasing displacements in the upper half of the mantle. It is therefore safe to conclude that the larger anomalies in tectonically active regions, as, for example, off the East Indies, the Philippines, and Japan, are connected with still stronger subcrustal currents throughout the mantle of the earth. The active source of these processes need not be close to the surface of the earth but may be deep seated. The location of the deep-focus earthquakes around the Pacific

Ocean has been interpreted as an indication that even at a depth of 700 km there is a difference in structure; such a hypothesis is not necessary, although it is possible. Forces acting at any given depth in the circum-Pacific region will set up stresses and plastic flow throughout the mantle. Wherever the stresses exceed the breaking strength of the material, earthquakes will accompany the processes.

As we have seen, defects of mass producing only relatively small gravity anomalies are able to produce noticeable plastic flow, if sufficient time is available. The lower limit, below which the processes stop or are too small to have observable effects, is of importance for the answer to the question whether certain small forces may produce considerable changes during long periods. Especially, is the *Polfluchtkraft* producing stresses with an order of magnitude of about one hundredth of the stresses in Scandinavia large enough to maintain plastic flow? If it is, then we should not be surprised to see continents being shifted by subcrustal flow during the history of the earth under the action of such small but persistent forces.

WORKS TO WHICH REFERENCE IS MADE

Antevs, E. (1939) *Late Quaternary upwarpings of northeastern North America,* Jour. Geol. vol. 47, p. 707-720.

Association d'Océanographie Physique (1939) *Monthly and annual mean heights of sea-level 1937,* Sécretariat de l'Association, The University, Liverpool 3.

——— (1940) *Monthly and annual mean heights of sea-level. Up to and including the year 1936,* Sécretariat de l'Association, The University, Liverpool 3.

Avers, H. G. (1931) *The determination of elevations,* in *Physics of the earth,* vol. 2, Nat. Res. Council, Bull. 78, p. 212-220.

Bell, R. (1896) *Proofs of the rising of land around Hudson Bay,* Am. Jour. Sci., 4th ser., vol. 1, p. 219-228.

Born, A. (1923) *Isostasie und Schweremessung,* Berlin.

Bowie, W. (1929) *Tilting of mean sea-level,* Gerlands Beitr. Geophysik, vol. 23, p. 97-98.

——— (1936) *Geodetic operations in the United States 1933-1935,* U. S. Coast Geodetic Survey, Spec. Publ., no. 207.

Burr, W. H., Hering, R., and Freeman, J. F. (1904) *Report of the commission on additional water supply for the city of New York,* New York, p. 650.

Chamber of Commerce of the District of Levis (1937) *An economic study and report of the St. Lawrence Ship Channel.*

Daly, R. A. (1925) *Pleistocene changes of level,* Am. Jour. Sci., 5th ser., vol. 10, p. 281-313.

——— (1934) *The changing world of the ice age,* Yale University Press, New Haven, 271 pages.

——— (1940) *Strength and structure of the earth,* Prentice-Hall, New York, 434 pages.

Dawson, W. B. (1917) *Tide levels and datum planes in eastern Canada,* Dept. Naval Service, Ottawa.

Egedal, J. (1933) *On the determination of the normal height of the sea-level round the Danish coasts,* Danish Meteorol. Inst., Publ.

Flint, R. F. (1940) *Late Quaternary changes of level in western and southern Newfoundland,* Geol. Soc. Am., Bull., vol. 51, p. 1757-1780.

Freeman, J. R. (1926) *Regulation of elevation and discharge of the Great Lakes,* Sanitary district of Chicago, 548 pages.

Gavelin, A. (1930) *Excursion from Boliden to Övertorneå,* Intern. Union Geod. Geophys., 4th Gen. Assembly, Stockholm, 1930. (According to Tsuboi, 1939; not in the proceedings).

Goranson, R. W. (1940) *Fracture and flow in stressed solids,* Am. Geophys. Union, Tr. 1940, pt. 2, p. 698-700.

Griggs, D. (1939) *Creep of rocks,* Jour. Geol., vol. 47, p. 225-251.

——— (1940) *Deformation of rocks in the laboratory,* Am. Geophys. Union, Tr. 1940, pt. 2, p. 159-162.

Gutenberg, B. (1933) *Tilting due to glacial melting,* Jour. Geol., vol. 41, p. 449-467.

——— (1939) *Viscosity, strength and internal friction,* in *Internal constitution of the earth,* New York, p. 361-384.

Haskell, N. A. (1935) *The motion of a viscous fluid under a surface load,* Physics, vol. 6, p. 265-269.

——— (1936) *The motion of a viscous fluid under a surface load. Part II,* Physics, vol. 7, p. 56-61.

——— (1937) *The viscosity of the asthenosphere,* Am. Jour. Sci., vol. 38, p. 22-28.

Heiskanen, W. (1939) *Catalogue of the isostatically reduced gravity stations,* Isostatic Institute, Intern. Assoc. Geod. Publ. no. 5, Helsinki.

——— (1940) *On the isostatic equilibrium of the earth's crust,* Sixth Pacific Sc. Congr., Proc., vol. 1, p. 165-173.

Jamieson, T. F. (1865) *On the history of the last geological changes in Scotland,* Geol. Soc., London Quart. Jour., vol. 21, p. 178.

——— (1882) *On the cause of the depression and re-elevation of the land during the glacial period,* Geol. Mag., vol. 9, p. 400-407, 457-465.

Johnson, D. (1929) *Studies of mean sea-level,* Nat. Res. Council, Bull. 70.

Johnston, W. A. (1939) *Recent changes of the land relative to sea level,* Am. Jour. Sci., vol. 237, p. 94-98.

Kuenen, Ph. D. (1940) *Causes of eustatic movements,* 6th Pacific Sc. Congr., Proc., vol. 2, p. 833-837.

Kukkamäki, T. J. (1939) *Über zwei dem Präzisionsnivellement sich anschliessende Fragen,* Veröff. Finnisch. Geodät. Institut, no. 26, Helsinki, p. 120-125.

La Cour, D. (1931) *Über Bestimmungen von Höhendifferenzen mittels Pegelbeobachtungen in Dänemark,* C. R., Comm. Géodésique Baltique, ciquième séance, Copenhague 1930, p. 221-233.

LaFond, E. (1939) *Variations of sea level on the Pacific coast of the United States,* Sears Foundation Jour. of Marine Research, vol. 2, p. 17-29.

Lidén, R. (1938) *Den senkvartära strandförskjutningens förlopp och kronologi i Angermanland,* Geol. Fören. Förhandl., vol. 60, H. 3, p. 397-404.

Marmer, H. A. (1931) *Mean sea-level,* in *Physics of the earth,* vol. 2, Nat. Res. Council, Bull. 78, p. 51-67.

Meinesz, F. A. V. (1937) *The determination of the earth's plasticity from the post-glacial uplift of Scandinavia; isostatic adjustment,* Acad. Amsterdam, Pr., vol. 40, p. 654-662.

Meissner, O. (1924) *Säkulare Schwankungen des Ostseemittelwassers,* Annalen der Hydrographie, vol. 52, p. 121-124.

——— (1926) *Jahrzehnte-Mittel der Konstanten des jährlichen Ganges des Wasserstandes einer Anzahl deutscher Ostseestationen,* Annalen der Hydrographie, Kleine Mitteilungen, p. 135-137.

Nansen, F. (1928) *The earth's crust, its surface forms, and isostatic adjustment,* Norske Videnskaps-Acad., Avhandl., Oslo, Kl. I, no. 12.

Niskanen, E. (1939) *On the upheaval of land in Fennoscandia,* Isostatic Inst. Intern. Assoc. Geod. Publ. no. 6, Helsinki.

Sauramo, M. (1939) *The mode of the land upheaval in Fennoscandia during Late-Quaternary time,* Société Géol. de Finlande, C. R., no. 13.

Schmidt, M. (1918) *Untersuchung von Höhen- und Lagenänderungen im bayerischen Alpenvorland,* Bayr. Akad. Wissensch., Sitzungsber., p. 373.

——— (1922) *Neuzeitliche Erdkrustenbewegungen in Frankreich,* Bayr. Akad. Wissensch., Math.-Phys. Kl., Sitzungeberichte, p. 1.

Scholaski, J. (1929) *Niveaux moyens comparés de la Mer Blanche, de la Mer Noire, et de l'Ocean Pacifique,* Acad. Sci. Paris, C. R., vol. 189, p. 24-25.

Schwinner, R. (1933) *Gebirgsbewegungen und Erdmessung in Süddeutschland,* Zeitschr. d. Deutschen Geol. Ges., vol. 85, p. 189-213.

——— (1935) *Gebirgsbewegungen im Südbayrischen Dreiecksnetz?* Zeitschr. d. Deutschen Geol. Ges., vol. 87, p. 59-60.

Stanley, G. M. (1939) *Raised beaches on east coast of James and Hudson Bays* (Abstract). Geol. Soc. Am., Bull., vol. 50, p. 1936-1937.

Tanner, V. (1930) *Studier över kvartärsystemet i Fennoskandias nordliga delar,* Comm. Géol. de Finlande, Bull. 88, Helsinki, 549 pages; with French summary.

Taylor, F. B. (1926) *The present and recent rate of land-tilting in the region of the Great Lakes,* Mich. Acad. Sci., Papers, vol. 7, p. 145-157.

Tsuboi, Ch. (1939) *Deformations of the earth's crust as disclosed by geodetic measurements,* Ergebnisse der kosmischen Physik, vol. 4, p. 106-168.

——— (1940) *Isostasy and maximum earthquake energy,* Acad. Tokyo, Proc., vol. 16, p. 449-454.

Tyrrell, J. B. (1896) *Is the land around the Hudson Bay rising?,* Am. Jour. Sci., 4th ser., vol. 2, p. 200-205.

van Bemmelen, R. W., and Berlage, H. P. (1935) *Versuch einer mathematischen Behandlung geotektonischer Probleme unter besonderer Berücksichtigung der Undationstheorie,* Gerlands Beitr. Geophys., vol. 43, p. 19-55.

Vignal, J., and Taton, R. (1935) *Rapport sur les nivellements de précision,* Traveaux de l'Association de Géodésie, tome 12, Paris.

Witting, R. (1918) *Hafsytan, Geoidytan och landhöjningen,* Fennia 39, no. 5, Helsingfors, 346 pages; with German summary.

——— (1922) *Le soulèvement récent de la Fennoscandie,* Geografiska Annaler, p. 458-487.

Woldstedt, P. (1929) *Das Eiszeitalter, Stuttgart.* (With about 800 references.)

Woodworth, R. W. (1927) *Tides in southeast Alaska,* U. S. Coast Geod. Survey, Spec. Publ., no. 127, p. 3-76.

California Institute of Technology, Pasadena, California.
Manuscript received by the Secretary of the Society, November 12, 1940.
Balch Graduate School of the Geological Sciences, Contribution No. 306.